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BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 71, PP. 1547–1574, 8 FIGS. NOVEMBER 1960

POST-PLIOCENE UPLIFT OF THE SIERRA NEVADA, CALIFORNIA

BY F. S. HUDSON

ABSTRACT

Matthes determined the position of three ancestral stages of the upper Merced River by projecting the gradients of hanging-valley tributaries to points above the present main stream. From the gradients of the two older stages he concluded that the Sierra Nevada range had been tilted westward, resulting in 9000 feet of uplift of the range crest at the head of the river, since the establishment in late Miocene time of the oldest, broad-valley, stage. Axelrod's work, based on paleobotany, indicates that at Granite Chief, on the summit of the range, about 100 miles north of the Merced River, the uplift has been about 5505 feet since late Miocene time.

In a redetermination of the old Merced channels by Matthes' method, the broad-valley stream is traced to the western edge of the range and thence into the ancestral San Joaquin Valley, by way of a pediment. A new calculation of the amount of tilting undergone by the range since broad-valley time, now considered early Pleistocene, is based on the assumption that certain well-graded stretches of the ancient stream had pretilt slopes equal to those of subjacent, well-graded parts of the modern river. The amount of tilting in the pediment region is determined by conventional geologic methods. The calculations indicate that the summit of the range, at the head of the Merced River, has been raised only 3930 feet since broad-valley (= early Pleistocene) time. The same methods applied to the Middle Fork of American River indicate that the contemporaneous uplift at Granite Chief was 2054 feet. Results of calculations made by other physical methods indicate the uplift at this place to have been between 2289 and 2745 feet since middle Eocene time and 1946 feet since the early Pliocene.

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Introduction

The Sierra Nevada is a north-northwest-trending range, 400 miles long and 55–75 miles wide, which separates the Central Valley of California, on the west, from the Great Basin on the east. It is highly asymmetric, and the rate of descent on its eastern side is 3–5 times that on its western slope. The altitudes of the peaks along the main crest increase progressively from 7000 feet, at the north end of the range, to more than 14,000 feet in the Mount Whitney region, 300 miles to the south, beyond which they decline to 6000 feet in the next 100 miles. In the areas of particular interest in the present paper, the highest peaks at the head of Merced River reach about 13,000 feet above sea level; maximum elevations at the head of Middle Fork of American River are about 9000 feet.

The rocks of the range fall readily into two major divisions. The older of these, the Bedrock complex, consists of metamorphosed sedimentary and volcanic rocks, of Paleozoic, Triassic and Jurassic age, intruded by plutonic rocks which range in character from ultrabasic to granitic, with the more acidic types predominant. The younger division, the Superjacent series, includes middle Eocene stream gravels and upper Tertiary rhyolitic, andesitic, and basaltic deposits, which rest on the bedrock with strong unconformity. The andesitic deposits, the thickest and most widespread part of this younger series, have been named the Mehrten formation and determined, from paleontologic evidence, to be early and middle Pliocene in age.

In the latitude of Merced River the Superjacent deposits are found only in the foothill region, but, beginning about 15 miles to the north, they are an increasingly important element of the Sierra proper in the form of a dissected cover of the bedrock on the west flank of the range, with dip near that of the topographic slope.

The manner and amount of uplift of the Sierra Nevada have long been matters of study and dispute. The early interest was stimulated by mining activity, but the shape and great length and height of the range still excite inquiry, despite the cessation of important mine working. The first worthwhile published contribution on this subject was that of Le Conte (1889), who held that, at the end of the Tertiary, the entire region from the Sierra Nevada to the Wasatch Range was bowed up as a huge arch, which collapsed into a series of basins and ranges and left the abutments (the Sierra and Wasatch scarps) as the boundaries of the Great Basin. The idea of great post-Tertiary upbowing may not be acceptable today, but geologists generally agree that the Sierra is an enormous block of the earth's crust which was tilted westward as its eastern side was dislocated from the Great Basin by faulting.

The presence of major faulting east of the range crest was confirmed by Lindgren (1897) and Turner (1897), and the westward tilting of the Sierra block was proved by Lindgren (1900) from his study of the gradients of one of the Tertiary auriferous channels of the western slope. Lindgren and Turner thought that the faulting occurred before the deposition of the andesite, but Louderback (1924) showed that postbasaltic major normal faulting took place along the range front, northwest of Lake Tahoe.

The results of recent work indicate that the deformation in this region was not solely due to normal faulting. Thus, 30 miles north of Lake Tahoe, the fault which separates Sierra Valley from the east foot of the range was thought to be "normal," but Durrell (1950) showed that the movement along it included more strike slip than dip slip. Hudson (1951) showed that, in the Donner Pass area, the range underwent several episodes of compressional deformation, including high-angle reverse faulting, beginning in middle Miocene and ending in early Quaternary time, before normal faulting.

Following the major part of this deformation, the higher parts of the range were glaciated. Slight dislocations seen in glacial deposits and earthquakes with epicenters along the range front indicate that faulting has continued.

Ten of the modern streams on the west slope of the Sierra are of major importance, in that

they drain the summit region. Their general course is southwest, at a right angle to the trend of the range, but only one, the North Fork of Stanislaus River, seems to be consequent on the tilting. The sinuous courses of the others preclude a consequent origin. The streams of interest in the present study are the Merced and Middle Fork and main American rivers. The former is near the longitudinal center of the range, and the latter is 100 miles to the north.

The courses and present gradients of middle Eocene and late Miocene streams on the west flank of the Sierra can be established by geological means, and the data of preglacial, Quaternary channels can be learned from physiographic evidence. The only ones of these ancient streams that conform closely in course with modern rivers are the Pleistocene predecessors of the Merced and Middle Fork of American rivers.

Calculation of the amount of uplift of the Sierra was first essayed by Lindgren. His work, based on the gradients of the Tertiary South Fork of Yuba River, led to the conclusion that the tilting of the range, since the cessation of the South Fork's erosive activity, was 60 feet per mile to the west-southwest. This calls for an uplift of about 3300 feet at Donner Pass on the summit of the range.

Matthes' (1930) work on the upper reaches of Merced River showed that the hanging-valley tributaries to that stream are of preglacial origin and that, by projecting their gentler, upstream gradients over the Merced canyon, three preglacial channels of the main river could be determined. He estimated that the uplift of the crest of the Sierra at the head of the Merced has been 9000 feet, since the inception of the tilting that caused the oldest of these channels to be erased by downcutting. His dating of this stream was late Miocene, but the writer believes that it is of early Pleistocene age.

Recalculating from Lindgren's data and additional information, Hudson (1955) found the post-middle Eocene uplift at Donner Pass to be a minimum of 1966 and a maximum of 2442 feet.

The latest estimate of this uplift is Axelrod's (1957). From the study of fossil plant remains found near the base of the Mehrten formation in the vicinity of Carson Pass, he concluded that the post-upper Miocene uplift was 6500 feet there and 5300 feet at Donner Pass.

The main purpose of the present work is to make a recalculation of the post-early Pleistocene uplift along the Merced River, based on a restoration of the earliest Pleistocene channel by Matthes' method, and to carry out similar work along the Middle Fork of American River.

Merced River

Modern Stream

From its headwaters a short distance south of Mount Lyell, a 13,090-foot peak on the crest of the Sierra Nevada, the Merced River flows 145 miles west-southwest to join the San Joaquin River on the west side of the Central Valley of California, at an elevation of 60 feet above sea level (Fig. 1). Its course is rather tortuous; the direct distance from head to mouth is only two-thirds of that along the channel.

The last 10 miles of the channel is on recent valley fill. East of this the stream crosses a north-northwest-trending belt of nonmarine Pleistocene and Tertiary sediments, 30 miles in width. There the low-water channel meanders on a flood plain, on either side of which are terraces and pediments carved on the Tertiary deposits by streams ancestral to the modern river.

East of the pediment belt, the 60-mile western slope of the Sierra Nevada traversed by the river consists of two strips of equal width: the eastern is underlain by granitic rocks, the western by metamorphic rocks. The latter is composed of two belts: Paleozoic on the east and Mesozoic on the west. These foliated rocks strike north-northwest, and most dip steeply east.

In its 64-mile course through the crystalline rock terrane, from Yosemite Valley to the pediment belt, the river flows in a narrow-bottomed, steep-walled canyon. In the greater part of this course the canyon is V-shaped in cross section, with wall slopes ranging from 40:100 to 75:100, but in Merced Gorge, which extends for 5 miles below Yosemite Valley, the cross profiles are U-shaped, with the upper parts of the walls sloping as much as 250:100. The uplands on either side of the canyon are rather rugged surfaces; the local relief increases from 450 feet in the western half of the range slope to more than 1000 feet on either side of the gorge. These uplands are an ancient land surface, which is being dissected by tributaries that have been rejuvenated by recent deepening of the canyon of the main stream, as shown by: (1) on the lower slopes of the range, all but

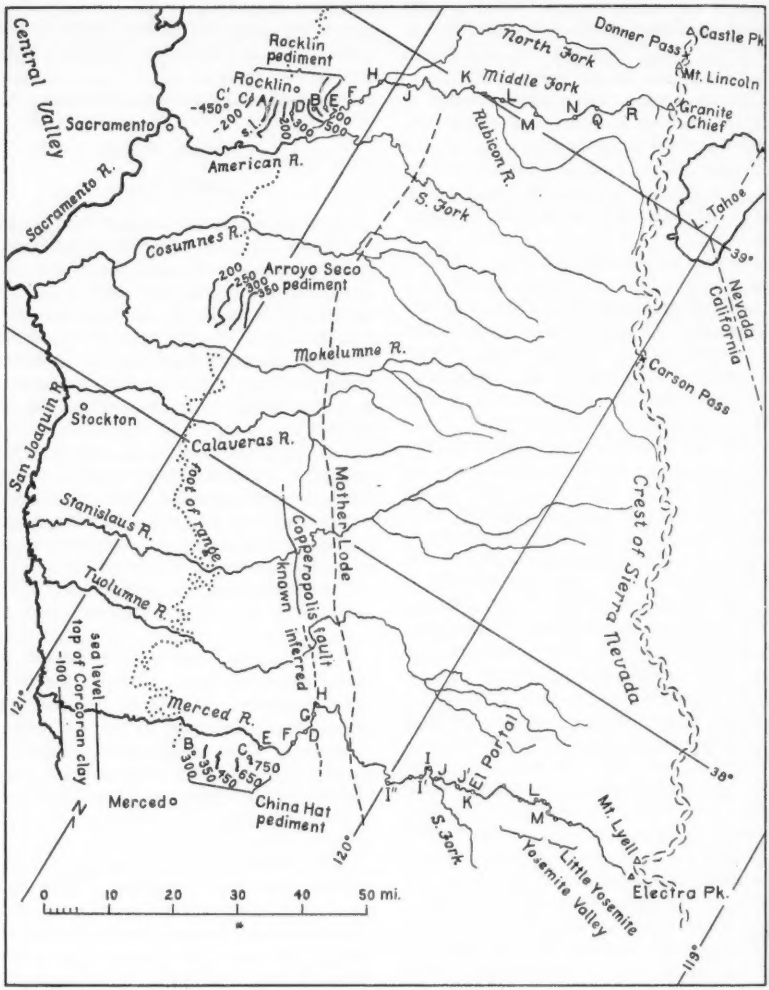

FIGURE 1.—KEY MAP
West slope of Sierra Nevada from American River south to Merced River

the most resistant rocks of the uplands are deeply decayed, and the more gently inclined surfaces are thickly mantled with red soil; (2) the boundary between canyon and upland is a definite brink; (3) the gradients of the typical upland streams decrease steadily from headwaters to points at, or some distance above, the brink, where they increase abruptly.

One who wants a clear idea of the remarkable topographic features of Yosemite Valley should consult Matthes (1930). The needs of the present paper will be served by a brief description of these features. The Merced River, from its source near 12,000 feet to Yosemite Valley, descends 7500 feet in 20 miles through a series of steep drops, separated by short reaches of gentle gradient. (*See* Figure 2.) The most important of the flat stretches is the westernmost, Little Yosemite Valley, and the 2100-foot drop from it to Yosemite Valley is the most notable

of the steep reaches, as within it are the 594-foot and 317-foot Nevada Falls and Vernal Falls. The canyons of the main stream and the valley are U-shaped, with the slopes of the upper walls in many places as much as 600:100 and exceptionally vertical. The low points of

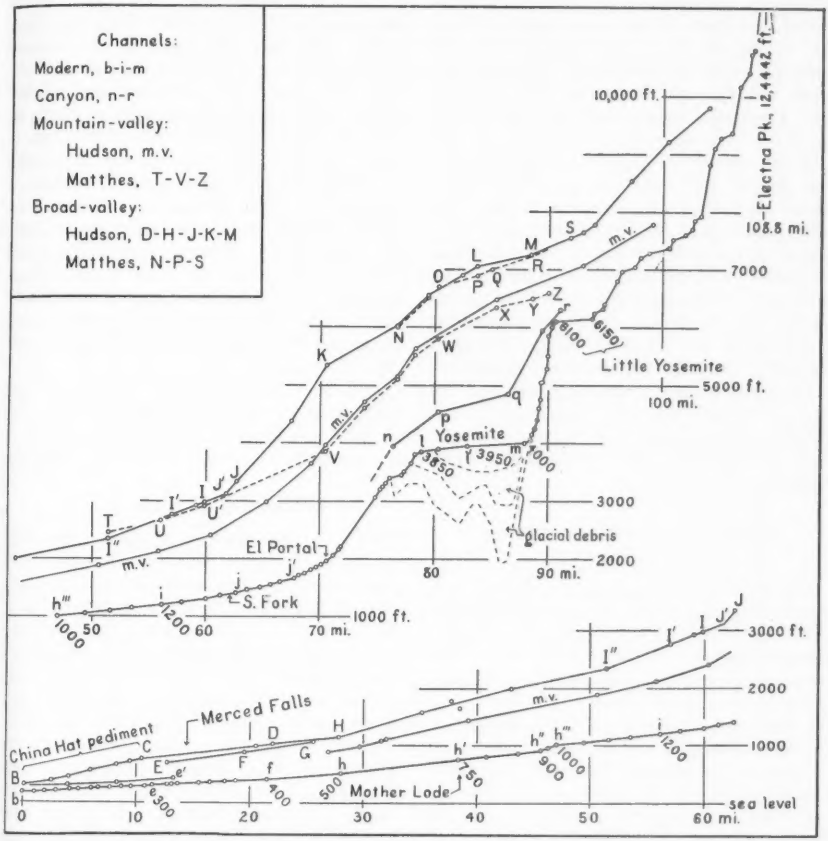

FIGURE 2.—LONG PROFILES OF MERCED RIVER CHANNELS

its principal tributaries, from the headwaters to the top of Nevada Falls, are U-shaped, with the upper slopes of the walls ranging from 50:100 to vertical. The low areas on the uplands on either side of the canyon are 1400–2200 feet above the river, and the peaks rise 800–1100 feet higher.

Below Vernal Falls is the Yosemite Valley, through which, on the greater part of a 12-mile course, the river meanders on a flood plain from a quarter to three quarters of a mile wide. On either side rise the spectacular cliffs that have made the valley famous. The cross profiles of the adjacent uplands are 3000–3750 feet above the valley, and the peaks are 400–800 feet higher.

Glaciation

Matthes (1930, p. 2) concluded that three successive glaciations followed the development of a canyon by the preglacial Merced River. These were called by him the Glacier Point, El Portal, and Wisconsin stages, and he believed that they deepened Yosemite Valley 1500 feet and widened it 1800 feet in the area of most intense action. Blackwelder (1931, p. 909) and

Birman (1954) concurred with Matthes as to the two later stages but believed that the Glacier Point stage had not been established. From a seismological survey made in 1935, Gutenberg, Buwalda, and Sharp (1956) concluded that three distinct stages occurred and that during the earliest of these the bedrock under the valley was excavated to a maximum depth of 2000 feet below the present valley floor. The base of the glacier of this stage flowed uphill over a granite sill, 8 miles west of, and 1400 feet above, the place of deepest excavation. Glacial debris and interglacial and postglacial delta and lake deposits brought the valley's surface to its present level. Long profiles of the channels cut during the three stages are shown in Figure 2, from m westward. The evidence of U cross profiles shows that the ice stream of at least one of the stages extended 4 miles from the sill down Merced Gorge to the 2300-foot level, 1½ miles above El Portal.

Preglacial is used in this paper to mean before the oldest recognized glaciation in the Yosemite region, believed to be much younger than the earliest continental glaciation of North America.

Tertiary and Later Sediments of the Lower Merced

The Ione sediments of Middle Eocene age, exposed immediately west of the Mesozoic rocks of the Sierra proper, and the overlying Valley Springs formation of Oligocene or early Miocene age, furnish no information pertinent to the present study. Some of the features of the succeeding deposits are of prime importance to the solution of the uplift problem. The oldest of these is the Mehrten formation, composed almost exclusively of andesitic detritus. Its age, based on vertebrate fossils, is early and middle Pliocene, according to Stirton and Goeriz (1942). The top of the Mehrten (Fig. 3B) is derived from Davis and Hall's (1959) Figure 3.

The Turlock Lake formation is described by Davis and Hall (1959, p. 10–16) as a sequence of gravel, sand, and silt layers, largely from nonvolcanic sources, with granitic materials predominant in the finer detritus. In the upper part of the formation is a layer of red clay, 3–8 feet thick, which they think is probably a buried soil. The Turlock Lake rests on the Mehrten with seeming conformity and is believed to have been deposited in late Pliocene and early Pleistocene time. The representation of the red clay and the bedding of the subjacent strata (Fig. 3B) is based on Davis and Hall's Plate 4.

Davis and Hall (1959, p. 16–20) state that their next younger formation, the Riverbank, overlies the Turlock Lake on a contact which has not been located precisely. They note that, in the subsurface, the beds of the younger formation have a gentler dip than those of the older. The dips above and below the red clay appear to indicate that this member marks an unconformity (Fig. 3B). The Riverbank is similar in mineral content to the underlying formation but is of finer grain. It contains a persistent member of blue clay, believed by Davis and Hall to be the equivalent of Frink and Kues' (1954) Corcoran clay of the region south of Merced River. The latter authors suggested a correlation of the Corcoran with the El Portal glaciation of the Yosemite region. Davis and Hall assigned the Riverbank to middle Pleistocene.

The Modesto formation, considered late Pleistocene by Davis and Hall (1959, p. 22) overlies the Riverbank unconformably on a contact marked by a layer of red clay (Fig. 3B).

Pediments of the Lower Merced

R. J. Arkley, College of Agriculture, University of California, has made, in addition to a soil survey, a study of the old land surfaces, exposed and buried, in the lower Merced area. His work on exposed soils is in print (1954), but his ideas on old erosion surfaces have not been published. The writer is indebted to him for the idea that pediments occur on both sides of the river in the Tertiary-Pleistocene belt. The one on the north, named herewith North Merced pediment, extends 18 miles west from e′ (Fig. 2), from 420 to 220 feet altitude, with gradient decreasing from 22 to 5 feet per mile. It cuts the Ione, Valley Springs, and Mehrten formations, and its surface is mantled by pebbles and cobbles of ancient Sierran types of rock. North Merced pediment is thought to be of the same age as the Arroyo Seco pediment, identified by Piper *et al.* (1939, p. 53) in the region 40–70 miles north of Merced River, and considered by him to have been formed after the last important uplift of the Sierra Nevada.

South of the river is China Hat pediment, named after the truncated, conical hill whose top is a residual outlier of the old erosional surface. It extends (C–B, Figs. 1, 2, 3) 9 miles west from its easternmost preserved part at the 750-foot level, to its disappearance under young sediments at 300 feet altitude. Its profile is

FIGURE 3.—PLEISTOCENE CHANNELS OF MERCED RIVER PROJECTED ONTO VERTICAL PLANE TRENDING N. 70° E.

sigmoid, and the slope is 45 feet per mile from C (750 feet elevation) to the 650-foot level, increasing to 68 feet per mile on approaching C″ (375 feet altitude), and then decreasing, through 50 feet per mile from C″ to C‴, to 28 feet per mile from C‴ to B. When formed, the pediment must have cut across the Ione and Valley Springs as well as Mehrten deposits, but the evidence of its course over the two older formations has been destroyed by erosion. This surface is strewn with material found in very few places in the Mehrten, including alluvial gold and boulders of pre-Upper Cretaceous rocks, which must have been brought to its present site after erosion had stripped a large part of the Sierra slope of its cover of Mehrten andesitic deposits. Accordingly, the pediment can be no older than late Pliocene.

This surface was undoubtedly formed by one or more ancestors of Merced River, and part of the post-Mehrten sediments found west of the pediment was transported over it during this cutting.

Davis and Hall indicated (Personal communication) their belief that the pediment is of the same age as some part of their Riverbank-Turlock Lake sequence and is perhaps the correlative of the red-clay layer in the upper part of the latter formation. In Arkley's view the pediment is compound, consisting of China Hat pediment proper, C to C‴ (Fig. 3B), and a younger surface, C‴ to B. He thinks that the older surface continues underground with a slope near that of the exposed part.

Following is a proposed history which brings these ideas into harmony. The carving of the pediment and the deposition of the Turlock Lake formation were concurrent and were terminated by tilting which led to the unconformity at the top of that formation and caused the river to take a new route to the valley. The B–C‴ part of the pediment is a relic of that time, which, according to Davis and Hall, was late early Pleistocene.

Ancestral Channels of Merced River

Earlier ideas.—The streams of the Yosemite uplands have relatively gentle gradients to points near the brink of the canyon, where they steepen abruptly and send their waters to the river in torrents, with waterfalls in many places. These hanging valleys are the upstream remnants of streams which entered an early Merced River through gently sloping channels; this was recognized by early authors, some of whom attributed the disappearance of the tributaries' lower reaches to downdrop of the Yosemite Valley by faulting. When it came to be thought that the valley was formed by glacial erosion, Gannett (1901) and Johnson (1911) concluded that the upper reaches of the tributaries were left hanging because their ice streams did not cut downward so rapidly as the trunk glacier.

Matthes (1930, p. 34–43) showed that the hanging valleys are not confined to the glaciated region. Citing Saxon Gulch and Feliciana Creek (T, U, Figs. 2, 3A), he stated that, although these are not of the clean-cut, spectacular type prevalent in the Yosemite region, they are nevertheless true hanging valleys. He concluded that the form of all these valleys is due to the inability of the feeble side streams to keep up with the rapid downcutting of the main stream, which had been accelerated by westward tilting of the Sierra block, whereas the tributaries were handicapped not only by their lesser volume but also by the fact that, their courses being parallel to the tilt axis, they were not steepened by the tilting. By prolonging the long profiles of the hanging valleys to the axis of Merced canyon, he restored the destroyed, lower reaches of these tributaries and determined the post-tilt (present) levels of their junctions with the main stream. Figure 4 gives examples of this method. From these restored points of confluence of the tributaries with the main stream, he constructed three long profiles (his Plate 27) showing the present positions of three ancestral Merced rivers which he named, in order of decreasing antiquity, the broad-valley, mountain-valley, and canyon stages. His two older stages are shown in Figures 2 and 3A by the graphs N–S and T–Z.

New determination of the old channels.—Employing the hanging-valley method, supplemented by the use of interfluve shoulders, the writer made a redetermination of Matthes' three stages and extended the study up- and downstream from the Yosemite area, on the new detailed maps which were not available to him. The results are shown in profile in Figures 2 and 3A, where the graphs D–M, m.v., and n–r represent, respectively, the broad-valley, mountain-valley, and canyon channels. According to Matthes' map of glacial deposits (1930, Pl. 29), none of the features used to fix the broad-valley channel from M downstream has been glaciated, whereas some of the basic data of the mountain-valley and canyon profiles are from glaciated terrain. The two younger profiles are thus much less credible than that of broad-valley channel.

Matthes carried his broad-valley profile only as far downstream as Cascade Creek (Fig. 2, N). The new profile continues westward, through K and J, to I, I′, and I″. The three latter points are close to U′, U, and T, the westernmost stations on his mountain-valley profile. Therefore, the lower reaches of the revised broad-valley channel correspond to what would have been the lower part of his mountain-valley stream, had he carried his work farther west.

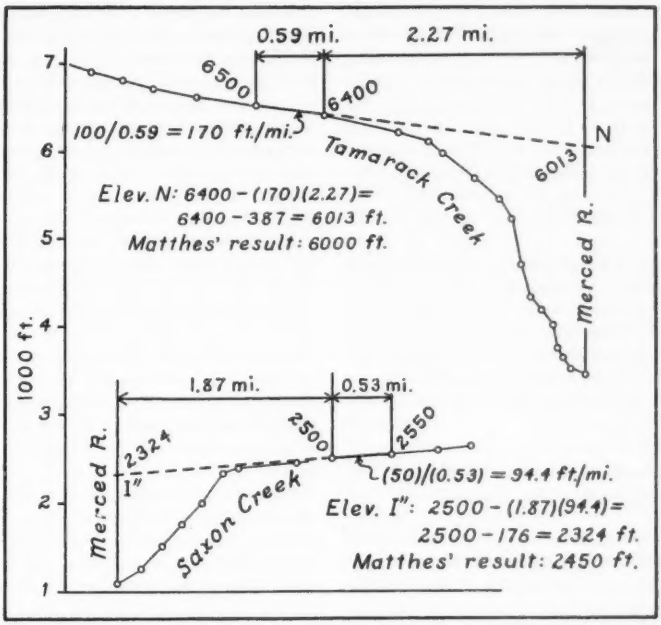

FIGURE 4.—EXAMPLES OF FIXING POINTS ON THE BROAD-VALLEY MERCED

Geologic age of the ancient channels.—Matthes thought that the broad-valley landscape was formed during the first half of the Tertiary, and that this period of erosion ended when the first of the andesitic deposits covered the western slope of the range. This was based on his correlation of the broad-valley surface with the surface preserved under the volcanic deposits of Tuolumne Table Mountain, 40 miles northwest of Yosemite Valley, and on the idea that the lower, andesitic part of these volcanic rocks is of later Miocene age. He believed that the mountain-valley stage began when the first strong tilting of the Sierra occurred in late Miocene time, that downcutting of the channel continued during the Pliocene, and that the canyon-stage cutting was caused by the major tilting of the range in early Quaternary time.

The writer accepts the idea that the three old channels were cut in preglacial time, and that two major tilting events occurred, but believes that these events occurred much later than Matthes thought. North of Merced River, for more than 100 miles along the middle and lower slopes of the west flank of the range, the uplands on either side of the main streams are the relics of an ancient landscape so similar to the broad-valley surface of the Merced region as to preclude doubt that they are parts of a single physiographic feature. As the Mehrten andesitic deposits are the subsoil of part of this landscape, the broad-valley surface is post-Mehrten, *i.e.*, no older than later Pliocene.

Referring to Figure 2, the stream which carved the C–C‴ part of the pediment could have been no younger than the broad-valley river, because the channel of a slightly younger stream, known from the fragmentary profile G–F–E, cannot be projected to C with a westward slope and must represent the river shortly after it abandoned the pediment. The cutting of the pediment by a pre-broad-valley stream

is unlikely. Any stream that cut both the buried and exposed parts of the pediment must have had a long and vigorous life and imprinted on the range strong evidence of its activity. Evidence of post-Mehrten, pre-broad-valley fluviatile action has not been found within the range. The writer concludes that the C–C''' part of the pediment was fashioned by the broad-valley stream during its latest activity. This correlation is indicated in Figures 2 and 3A by the lines joining D to C. The date of the final broad-valley channel is then that of the unconformity at the base of the Riverbank formation, which, following Davis and Hall, is the boundary between lower and middle Pleistocene.

The writer believes that the cutting of the mountain-valley channel and the deposition of the Riverbank sediments occurred concurrently, until terminated by the tilting which led to the post-Riverbank unconformity. Accordingly, in Davis and Hall's chronology, the end of the mountain-valley stage was in late middle Pleistocene time.

Deformation of the Broad-Valley and Mountain-Valley Channels

Matthes' conclusions.—Matthes' idea (1930' p. 24, 43–44) was that the Sierra Nevada range is a single block of the earth's crust which has been dislocated and tilted southwestward, apparently without appreciable internal deformation, in two successive movements, the earlier at the end of his broad-valley stage, the later following his mountain-valley stage. He cited the method used by Lindgren (1911, p. 46–48) to calculate the tilting of the Sierra from the gradients of the Eocene gold-bearing channels of the northern part of the range. This was based on the observation that the present slope of each stretch of the ancient channel depends upon the relation between its trend and that of the axis of tilt of the range; the reaches directed west-southwest were steepened most by the tilting, those extending east-northeast had their slopes decreased or even reversed, and those trending north-northwest or south-southeast still have their original gradients.

There must be one or more reaches of north-northwest or south-southeast trend in order to use this method. Matthes found only one such stretch, the north-northwest segment of his mountain-valley channel between the mouths of Illilouette and Indian creeks (Fig. 2, Y to X). A résumé of his calculation follows. The present elevations of Y and X are 6500 and 6350 feet. The length of course Y–X being 2.5 miles, its present gradient is (6500 − 6350)/(2.5) = 60 feet per mile, which, on the assumption that Y–X is parallel to the axis of tilting, was also its pretilt gradient. The southwest-trending reaches immediately above and below Y–X have gradients of 175[1] and 140 feet per mile, which, for the purposes of the calculation, he assumed to be 150 feet per mile, each. Assuming that these courses were at right angles to the tilt axis and that their pretilt gradients were the same as that of Y–X, he found the amount of the post-mountain-valley tilting to be 150 − 60 = 90 feet per mile.

Matthes was unable to establish a north-northwest course of broad-valley age here but offered an approximate calculation. He states that the present slope of this part of the broad-valley channel, indicated by his Plate 27A, is 114 feet per mile.[2] Deducting the 90 feet per mile of the later tilting left 24 feet per mile for the slope of the broad-valley channel, before that tilting. The gradient before the first tilting must have been less—he thought it was about 12 feet per mile.

To calculate the amount of uplift for each of these tiltings of the range required the determination of the positions of the hinge lines. This, Matthes said, was done by the extension of the broad-valley and mountain-valley profiles toward the foothills and the similar extension of the old profiles of the San Joaquin and northern rivers investigated by Lindgren (1911). On these bases he concluded that the crest of the Sierra at the head of Merced River was raised about 3000 feet in the uplift of the pre-mountain-valley stage and an additional 6000 feet in the second uplift. He did not include the calculations that gave these results.

Criticism of Matthes' work.—Reference to the newly determined profiles (Fig. 3) will show that, if the range was tilted as much as 90 feet per mile after the mountain-valley period, the pretilt gradients of the lower and middle reaches of both the broad- and mountain-valley channels must have been negative, *i.e.*, their streams flowed eastward. This result fully condemns Matthes' conclusions, provided the new profiles are accepted as sound.

Neither the method nor the results of his determination of the positions of the tilt axes were set out in Matthes' paper, but they can

[1] Matthes' profile indicates only 77 feet per mile.
[2] The amount indicated is actually 81 feet per mile.

be learned by simple calculations. His earlier tilting of 12 feet per mile is said to have raised the crest of the range 3000 feet. The axis for this tilt would be 3000/12 = 250 miles west of the crest. This is in the Pacific Ocean, 100 miles west of Monterey. This is implausible in itself and, when the full effects are considered, impossible. For instance, the western edge of the Sierra at Merced Falls, which is 59 miles from the crest of the range, measured along a line perpendicular to the tilt axis (*see* Figure 3A), would be elevated [(250 − 59)/(250)] [3000] = 2292 feet in such a tilting. He said that the summit of the range rose 6000 feet during the second tilting. This requires an axis of tilting situated 6000/90 = 67 miles from the crest. This later tilting would raise the western edge of the Sierra an additional (67 − 59)(90) = 720 feet, making the total post-broad-valley uplift, here, 3012 feet.

A mile west of Merced Falls, Mehrten deposits rest on Ione, middle Eocene, beds at an elevation about 300 feet above sea level. Matthes' tilts would have elevated this locality 3012 − 12 − 90 = 2910 feet, and, if his correlation of the base of the andesitic (Mehrten) deposits with the broad-valley stage of Merced River is correct, the pre-Mehrten surface here stood at 2910 − 300 = 2610 feet below sea level before the earlier tilting. The Mehrten is a continental formation, and, as Matthes said, its base is an ancient landscape. One might devise a complicated geologic history, involving sinking as well as uplift, to reconcile a 2910-foot uplift with the present 300-foot elevation at this place. But this would be of no avail. The 2910-foot uplift, based on Matthes' work, is net uplift—*i.e.*, the net effect of both positive and negative movements. The acceptance of this uplift figure would lead to the conclusion that the pre-Mehrten "land" here stood at 2610 feet below sea level when the deposition of the nonmarine Mehrten began.

If one rejects the correlation of the base of the andesitic deposits with the broad-valley stage and accepts the newly determined profile, which indicates the present elevation of the ancient channel to be 820 feet at a point 1 mile west of Merced Falls, the use of Matthes' tilt factors yields results with reduced arithmetical discrepancy but still absurd—a pretilt position of the broad-valley channel at 820 − 2910 = 2090 feet subsea.

Inconsistency between Matthes' tilt factors and his basic data of channel elevation is evident at several places on his profiles. Consider, for instance, the two westernmost reaches of his mountain-valley channel, the data for which are:

Stations	From Matthes' Plate 27		Difference in elevation (feet)	Distance on S. 70° W. proj. (miles)
	Elevation (feet)	Distance (miles)		
T	2450			
		4.6	200	3.8
U	2650			
		4.0	250	2.2
U'	2900			

Using his factor of 90 feet per mile for post-mountain-valley tilting, the original gradients of the two reaches would have been:

T–U: [200 − (90)(3.8)]/(4.6) = −31 *feet per mile*
U–U': [250 − (90)(2.2)]/(4.0) = 15.

The first result is absurd, the second implausibly low.

Another instance is that of his broad-valley reaches O–P and P–Q, which, on the S. 70° W. projection (Fig. 3A), have gradients of (200)/(2.4) = 83 and (100)/(1.1) = 93 feet per mile, respectively. To have such gradients after a westward tilt of 102 feet per mile requires negative pretilt gradients.

Calculation of Post-Broad-Valley Uplift

Matthes' idea (1930, p. 43) that the range was deformed by tilting to the west-southwest or, in a round number, S. 70° W., is accepted because this is the general direction of the surface slope on the western flank of the Sierra and is perpendicular to the grain of the pre-Tertiary rocks and to the strike of faults. Figures 3A and 3B are profiles with this trend, onto which the channel stations and other pertinent points are projected from N. 20° W. or S. 20° E.

The idea of tilting of a rigid block is not acceptable. The China Hat pediment cannot be a part of such a block. Its rate of tilting must have been less than the present slope (22 feet per mile) of the Turlock Lake sediments. As these and the Riverbank formation were probably deposited above the sea, the tilt axis was east of the present sea-level point of the top of the latter formation, which is about 90 miles, on the projection, from the

range summit. Assuming rigidity during tilting, the factor of < 22 feet per mile calls for a post-broad-valley uplift of the crest of the Sierra of less than (90) (22) = 1980 feet, an absurdly small amount when compared with the lofty escarpment on the eastern front of the range. If the C–D block was rigidly attached to the range to the east of D during the tilting, the result is even less reasonable.

Evidently, the tilting was accompanied by bending or faulting east of the C–D block. From Turner and Ransome's (1897) geologic map of the Sonora quadrangle, the post-Jurassic Copperopolis fault, shown on Taliaferro and Solari's (1948) map in the area 25 miles northwest of D, would, if it extended south-southeast to Merced River along the strike of the formations (Fig. 1), pass between D and H. The calculations indicate that the deformation east of C–D was along a single zone, passing through D, or a short distance east of that point. This suggests a late renewal of activity on a southern extension of the Copperopolis fault, but, for the purpose of the uplift estimates, the writer assumes that the deformation of the Pleistocene channels consisted of bending along a zone that passed through D.

The calculation of tilt in the pediment area was made by conventional geologic methods. The estimates of uplift within the range are based on the assumption that certain evenly graded stretches of the broad-valley channel had pretilt gradients equal to those of smooth, subjacent parts of the modern stream. This will seem unwarranted to those who may regard the present Merced as a poorly graded river and accept Matthes' idea (1930, p. 44) that the broad-valley channel was "well graded." On his profile the course between Illilouette and Bridal Veil creeks (Fig. 2, R–O) is remarkably smooth, but he fails to take into account the abrupt steepening at O, where the post-tilt downstream gradient increases from 59 to 194 feet per mile. This part of his broad-valley channel was surely not "well graded."

The modern channel appears to be no less "well graded" than the broad-valley channel; the writer believes that part of it is better graded than the ancient stream. Below Merced Falls the present stream, in its course across the Tertiary belt, is flanked by terraces whose gradients are essentially the same as adjacent parts of the river. This, according to Woodford (1951, p. 816), is a criterion for a graded stream. Figure 2 shows that above Merced Falls the gradient increases regularly to j′, 3 miles below El Portal, where the slope steepens abruptly from 50 to 108 feet per mile. The smooth concavity of the profile of this stretch of channel is marred only by a slight convexity at h‴, where the stream passes from Paleozoic rocks on the east to Mesozoic on the west. The irregularity in the profile here may be ascribed to a difference in the resistance to erosion of the two formations. The writer believes that the concave nick at j′, 1 mile downstream from the main granite-schist contact of the region is the result of a similar situation.

The writer regards the stretch of channel from j′ to Merced Falls as graded, because its smooth profile appears to be in mathematical continuity with the profile west of the latter point. This idea is not urged, however, since the objective is not to apply physiographic terms but to compare the stage of grading of the present stream with that of the broad-valley channel. A direct comparison of the two channels upstream from j′ is not possible because of the anomalies created by glacial erosion, but this difficulty can be overcome by the use of Matthes' canyon-stage stream as part of the basis of comparison. The graph n–p–q–r in Figure 2 is a new determination of the profile of this channel. It is close to that of Matthes, except as to point q, which is about 400 feet lower than the position shown by him (1930, Pl. 27 A). Matthes' idea (1930, p. 1) that the canyon stage immediately preceded the earliest glaciation seems acceptable. Therefore the profile of that stage is probably a fair approximation of what one would see today, had there been no ice age. Such a stream should have a strong convex nick at n, or a short way west of that point (Fig. 2). This nick has its analogue at K, 6 miles downstream on the broad-valley channel, and the concave nick at j′ has its counterpart at J′, a like distance downstream on the ancient river. The modern river has evidently gone much farther than the broad-valley stream in demolishing the Sierra.

The writer believes that the foregoing is adequate evidence for the conclusion that the modern Merced, from j′ westward, has attained a stage closer to one of equilibrium than that reached by the broad-valley river below J′. Accordingly, where it is assumed in the following calculations that certain reaches of the ancient channel had pretilt gradients equal to those of underlying stretches of the modern stream, the gradients so determined are likely to be lower than those that existed. This will tend to make the calculated amounts of uplift greater than the true values.

The Sierra, since broad-valley time, has

doubtless undergone many individual mountain-making movements. Calculations will be made of only the total of these movements, represented by the deformation of the broad-valley channel. It seems not worthwhile to attempt a calculation of the post-mountain-valley fraction of the uplift, because the profile on which it would be based is deficient in two respects: it cannot be carried downstream past the 880-foot level, and its upper reaches are based on hanging valleys that have been glaciated.

Lack of data at critical bends in the ancient channel precludes the determination of pretilt gradients and amount of uplift by the previously described Lindgren method or by the formulas applied (Hudson, 1955) to the Eocene Yuba River.

A TO C: The writer assumed that the axis of tilting was at the present sea-level position of the top of the Riverbank formation (A, Fig. 3B) and that the rate of tilt was equal to the slope of the Turlock Lake strata along the S. 70° W. profile, namely 22 feet per mile. As the distance A–C on that profile is 28.8 miles, the uplift of C with respect to A was (22) (28.8) = 634 feet.

C TO D: This stretch measures 11.5 miles along the channel and 7.7 miles on the S. 70° W. projection. The present rise from C to D is 253 feet. The evenly graded, subjacent reach e-f (Fig. 2) of the modern stream has the slope (100)/(10.1) = 9.9 feet per mile. This is assumed to have been the pretilt gradient of the overlying broad-valley channel. Accordingly, the original rise from C to D was (11.5) (9.9) = 114 feet, and the increased elevation of D vs. C was 253 − 114 = 139 feet. The rate of tilting was (139)/(7.7) = 18.1 feet per mile.

D TO H: The present rise from D to H is 135 feet in 5.8 miles along the channel but only 0.9 mile on the S. 70° W. projection. The underlying stretch, f–h, of the modern stream has the gradient (100)/(6.5) = 15.4 feet per mile. Assuming that this was the pretilt slope of D–H, the original rise from D to H was (5.8) (15.4) = 89 feet, and the increase in elevation of H with respect to D was 135 − 89 = 46 feet. This calls for tilting at the rate of (46)/(0.9) = 51.1 feet per mile.

H TO I: In this stretch the broad-valley channel rises 1832 feet in 32.0 miles along the gutter or 19.4 miles on the projection. Save for one short reach, h″ − h‴ (Fig. 2), the underlying segment, h–i, of today's river is evenly graded, with slopes ranging from 20.8 to 24.0 feet per mile. It is assumed that the average of these slopes, 22.3 feet per mile was the pretilt gradient of H–I. Accordingly, the original rise along this stretch was (22.3) (32.0) = 714 feet, and the increase in height of I with respect to H, caused by tilting, was 1832 − 714 = 1118 feet. The required rate of tilting is (1118)/(19.4) = 57.6 feet per mile.

I TO K: The slope changes abruptly from 220 to 80 feet per mile, at J′ (Fig. 2) on the broad-valley channel. One must go nearly 10 miles upstream, to the 2000-foot level, to find a nick of comparable severity on the modern river. This might suggest deformation at J′, but several considerations show that this is unlikely: (1) the subjacent terrain affords no evidence of late deformation, (2) no nick on the mountain-valley channel is comparable to that at J′ on the older channel, and (3) there are severe nicks at relatively low levels on the modern channels of other Sierra streams, such as that at 1200 feet elevation on the Middle Fork of American River (l, Fig. 5A) where the gradient changes from 132 to 38 feet per mile. The writer concludes that the J′ nick is a normal erosional feature and that no deformation occurred within the I–K block during the tilting. Accordingly the writer assumes that the I–K block was rigidly attached to the block to the west during the tilting and that the tilt fraction of 57.6 feet per mile found for H–I is applicable to I-K. As the length of the latter stretch on the projection is 7.1 miles, the uplift of K with respect to I was (7.1) (57.6) = 409 feet.

K TO THE SUMMIT OF SIERRA: The uplift of this part of the range can be calculated by the method used for the H–K segment—that is, by assuming that the gentle gradients in the l–m stretch of the present river are equal to the pretilt slopes of the equivalent part of the ancient stream. This assumption seems to be much less credible than that upon which the H–K calculation was based, because it would compare a channel in postglacial alluvium with one carved from granitic rock. Accordingly, the calculation will proceed from the assumption that the block east of K was rigidly attached to the H–K block during the deformation. The basis for this is that nothing is known in support of the contrary. The abrupt change in the slope of the broad-valley profile at K might be the result of bending or faulting, particularly as it is immediately upstream from the zone of faulting at the principal schist-granite contact of the region; however, the K feature appears to be a normal erosional nick, since it is less severe than its analogue on the modern stream,

and there is no evidence of post-Pliocene movement at the fault zone. The distance from K to the crest along the projection being 27.5 miles, the use of the H–K rate factor makes the uplift of crest vs. K (57.6)(27.5) = 1584 feet.

TABLE 1.—POST-BROAD-VALLEY UPLIFT ALONG MERCED RIVER

Stations	Projection distance (miles)	Uplift		
		Rate (feet per mile)	Amount (feet)	
			Along courses	Cumulative
A				..
	28.8	22.0	634	
C				634
	7.7	18.1	139	
D				773
	0.9	51.1	46	
H				819
	19.4	57.6	1118	
I				1937
	7.1	57.6	409	
K				2346
	27.5	57.6	1584	
Summit				3930
Totals	91.4	..	3930	..

Over-all average rate of tilting: 43.0 feet per mile

Essentially the same result may be obtained by the equal-gradient method. Along L–M, the reach of gentlest slope east of K, the rise is 197 feet in 4.7 miles by channel or 2.5 miles along the S. 70° W. projection. This reach is underlain by the modern stream's course l′–m, on which the slope is 10 feet per mile. Assuming that this was the pretilt gradient of L–M, the original rise from L to M was (10)(4.7) = 47 feet, and the increased height of M vs. L, caused by tilting, was 197 − 47 = 150 feet. The rate of tilt, then, was (150)/(2.5) = 60 feet per mile, and the uplift of the crest vs. K was (60)(27.5) = 1650 feet.

A TO THE SUMMIT: The summary of the foregoing results (Table 1) shows the post-broad-valley uplift of the range summit to have been 3930 feet. This is the total effect of two episodes of tilting which, in Davis and Hall's chronology, occurred at the end of the lower Pleistocene and in late middle Pleistocene time. The latter, the post-mountain-valley event, from the relative heights of the two Pleistocene channels above the modern stream (*see* Figure 2), appears to have been responsible for nearly two-thirds of the total uplift.

MIDDLE FORK AND MAIN AMERICAN RIVER

Present-Day Stream

The three major parts of the American River system are, in the order of lessening importance, the South, Middle, and North forks (Fig. 1). All rise near the summit of the Sierra Nevada and flow westerly. The latter two join near the western edge of the rugged part of the range to form the main stream, which, after crossing a belt of foothills, receives the flow of South Fork near the eastern border of the Central Valley. As attempts to calculate the uplift of the Sierra along North and South forks failed, whereas a solution was obtained from the data of Middle Fork, only the latter will be considered further.

From its source south of Granite Chief, a 9006-foot peak on the summit of the range (Fig. 1), Middle Fork, after a 5-mile westerly course, flows west-southwest for 55 miles to its junction with North Fork. Midway on this stretch, an important tributary, Rubicon River, enters from the southeast. Below the confluence with North Fork, the stream flows 22 miles south-southwest to its junction with South Fork and thence 30 miles west-southwest to Sacramento River.

The country traversed by the stream is divisible, geologically, into four north-northwest-trending belts of about equal width. In the western belt the channel meanders on a flood plain about 1 mile in width, which has been cut 20–100 feet below the surface of a mass of gently inclined Quaternary deposits. The adjacent belt to the east consists largely of metamorphosed Jurassic sediments and volcanic rocks. This is succeeded by a strip of metamorphosed Paleozoic deposits, which is followed by the summit belt in which granitic rocks predominate. The foliated rocks of the two middle belts generally strike north-northwest and dip steeply east.

The Middle Fork has many features similar to those of Merced River. The long profiles of the two streams (Figs. 2, 5A) are remarkably alike in form, being made up of two concave curves, joining at sharp, convex nicks. Both rivers flow in steep-walled canyons, outside of which are uplands with fairly strong local relief. The uplands adjacent to the Middle Fork exhibit the same evidence, item by item, from which the writer concluded that the Merced canyon is the result of recent incision into an ancient landscape.

Cross profiles of the Middle Fork canyon show that glaciation reached the 4250-foot

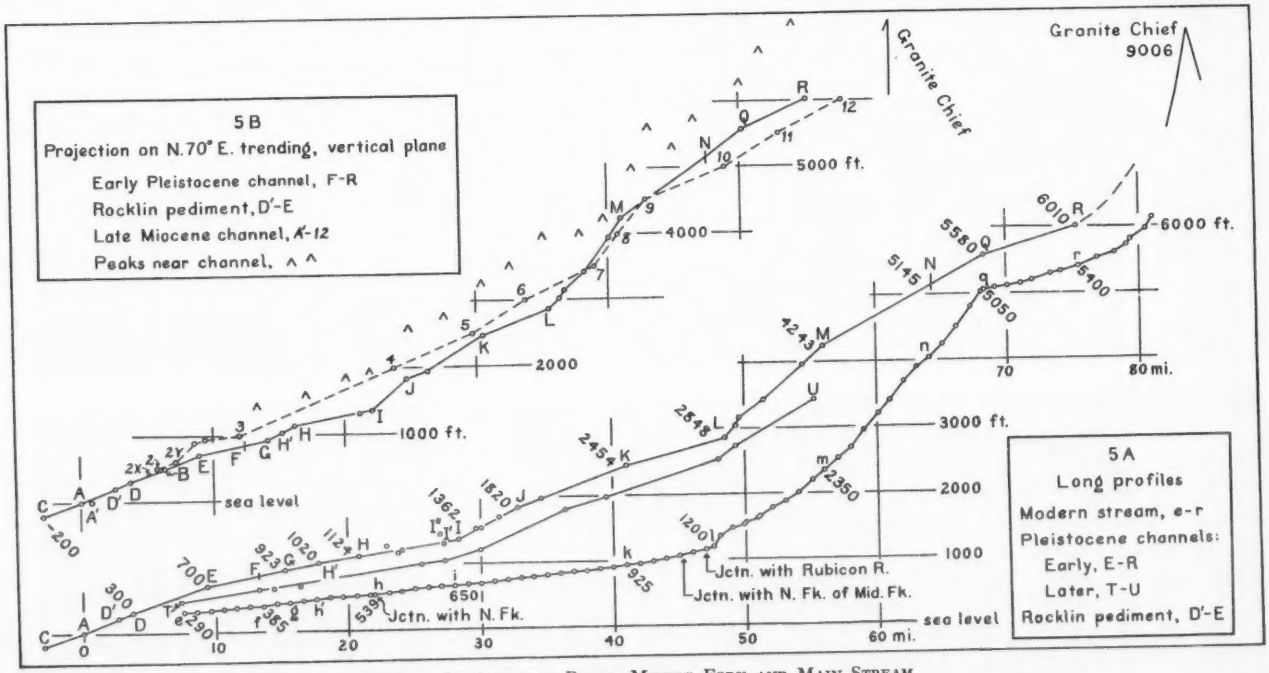

FIGURE 5.—AMERICAN RIVER, MIDDLE FORK AND MAIN STREAM

A. Long profiles of modern stream and Pleistocene channels. B. Projection of early Pleistocene and late Miocene channels on vertical plane trending N. 70° E.

level and that postglacial stream erosion has cut a V groove in the bottom of the U valley as far upstream as the 5050-foot level.

Pediments

Another resemblance between the Merced and American rivers is the presence of two pediments outside the canyon mouth of each stream. The older of the American pair is the Rocklin pediment (Fig. 5, D′–D–E), named after the town on its western edge. It lies west of the lower canyon stretch and is separated from it by a low ridge of pre-Upper Cretaceous rock. This pediment was carved on granite and Mehrten andesitic deposits. It was once covered by a sheet of gravel composed almost exclusively of detritus of pre-Upper Cretaceous rocks. Severe dissection has left only six isolated remnants of this sheet, but these furnish enough data for a determination of the pediment's original form, which, in plan, was convex toward the southwest. Its steepest slope is about 70 feet per mile from the 700-foot level (Fig. 5, E), S.48°W., to the place where it passes beneath young alluvium at 210 feet above sea level.

The features of this ancient erosion surface are so much like those of the Merced's China Hat pediment, that they must be the same age.

The dissection of the Rocklin pediment resulted in the development of a lower, unnamed pediment. As it furnishes no data pertinent to the problem of tilting of the range, it is not shown on the profiles.

Preglacial, Pleistocene Channels

Using the methods applied to the Merced River, the writer determined two ancient channels, an earlier one from R to F, and a later one from T to U (Fig. 5A). These are certainly preglacial and are believed to be of Pleistocene age for reasons similar to those cited in showing that the Merced's broad-valley and mountain-valley stages are of that age. The similarity in form of the long profiles of the ancient channels of the two rivers, together with the presumption that these channels were cut during two periods of orogenic quiescence, which prevailed throughout the Merced-American segment of the range, suggest strongly the contemporaneity of the American stages with Matthes' broad- and mountain-valley stages. However, the terms Early American and Later American are used here because precise correlation is not possible.

As was the case with the profile of the mountain-valley stage, that of the Later American appears to be of no use in calculations of the uplift of the Sierra and will not be considered further. Referring to the long profile of the Early American (Fig. 5A), the writer determined stations R to G by the hanging-valley method, but, farther downstream, no other points could be fixed by that procedure. Two miles below G, at f, the course of the modern river turns abruptly from S. 57° W. to S. 13° W. The writer believes that the ancient stream made no such turn but, after following the course G–F, superjacent to g–f on the modern channel, continued southwest to the head of the Rocklin pediment, crossing the site of what is now the bedrock ridge which flanks the present river on the west. The only data from which to depict this course are channel points G and F, and station E, on the highest preserved part of the pediment. Accordingly, for the purposes of the calculation, the writer assumes that the Early American followed a straight S. 48° W. line from F to E, which crosses the ridge 120 feet above a saddle where the elevation is 640 feet above sea level. The idea that the ancient stream had a course much like this is substantiated by the following:

(1) As may be seen on the long profile (Fig. 5A), the E–G course has a gradient close to that of G–H′, the next course upstream, and its characteristics conform well with those of the previously determined features of the Early American channel.

(2) The pediment must have been cut by a powerful stream, and the Early American was the only river available for this role.

(3) That a river actually crossed the site of the present ridge is attested to by a remnant of stream gravel on the ridge crest 2000 feet southeast of the 640-foot saddle, at 620–640 feet elevation. This gravel consists of shingle and rounded pebbles and boulders of a variety of pre-Upper Cretaceous, Sierran rocks, including gneissic diorite, gabbro, and quartz. It is found in a sandy soil, resting on granite. The pebbles of the shingle are from less than 1 inch to 4 inches in major diameter, and their thicknesses are from one-fourth to one-sixth of that diameter. The rounded constituents are from 1½ to 16 inches in major diameter. This material has no resemblance to any known Teritary gravel of the region. On the basis of composition alone, it might be Cretaceous, but its lack of cementation precludes such a correlation. Accordingly, the writer believes that it is Pleistocene, and, if this is correct, its position indicates that it was deposited by a stage

of the river slightly younger than Early American.

If a river of this date crossed the ridge site, the older stages of the stream must have done likewise.

Calculation of Post-Early American Uplift Along Middle Fork and Main American River

To calculate post-Early American uplift along Middle Fork and main American River the writer used methods and assumptions similar to those used on the ancient Merced, except that, in the summit region, the tilting is assumed to have been toward N. 54° W., instead of S. 70° W. A comparison of the profiles of the modern stream shows that the e–l section of the American is as near equilibrium as the e–j course of the Merced. The overlying segments of the Early American and broad-valley channels, E–L (Fig. 5A) and C–J′ (Fig. 2), do not match so well; the latter are better graded. Accordingly when, as is done in some of the calculations, certain parts of the E–L stretch of the Early American are assumed to have had the same gradients, before tilting, as those of underlying sections of the modern stream, the error is likely to increase the calculated amount of uplift.

The slope of the modern channel was also used, though with less confidence, as the basis of estimates of tilting in the summit region. The tilt of Rocklin pediment was calculated by assuming that its nose had a pretilt slope equal to that of a comparable part of the Merced's China Hat pediment.

The data used in the calculations are shown in long profile in Figure 5A, and their projection on a vertical plane trending S. 70° W. is shown in Figure 5B.

TILTING OF ROCKLIN PEDIMENT: The steepest part of China Hat pediment drops 175 feet, from C′ to C″ (Fig. 3), in a distance of 2.5 miles along a S. 70° W. line. The present gradient of this stretch is (175)/(2.5) = 70.0 feet per mile and, as the rate of tilt was found to have been 22 feet per mile toward S. 70° W., the pretilt gradient of C′–C″ was 70 − 22 = 48 feet per mile. The writer assumes that this was also the original slope of the comparable part of the Rocklin pediment (E–D, Fig. 5B), where the present decline is 400 feet in 5.6 miles along a S. 48° W. line. This assumption seems reasonable because of the similarity of the canyon profile of the Early American to that of the broad-valley Merced.

Using the 48 factor, the pretilt rise from D to E on the Rocklin pediment was (48.0)(5.6) = 269 feet, and the increase in elevation of E with respect to D, caused by tilting, was 400 − 269 = 131 feet. As the distance E–D on the S. 70° W. projection is 5.2 miles, the rate of tilting in that direction was (131)/(5.2) = 25.2 feet per mile.

As good marker formations have not been recognized in the Pleistocene deposits under the valley, west of Rocklin, the fixing of the position of the tilt axis of Rocklin pediment cannot be done with the same exactness as that of the China Hat pediment. However, the writer believes that this determination can be made with a precision consistent with that of the other calculations of uplift along the ancient American. A trigonometric solution, based on the slopes of various parts of the pediment, placed the tilt axis at B, 2 miles west of the highest preserved part of the pediment. This result is based on crude and meager data and seems incredibly far mountainward in comparison with what is thought to be the well-established axis of China Hat pediment. Therefore it is rejected in favor of the product of a simpler method.

The pediment surface on the Mehrten deposits passes under valley alluvium at D′, at 215 feet above sea level (Fig. 5B). The pediment cannot be identified, as such, west of this point, but the top of the Mehrten is recognizable in the logs[3] of water wells, drilled at such locations that the subsurface position of this horizon can be determined accurately from 100 feet above sea level through sea level at A, to 500 feet below sea level. West of D′, the horizon of pediment age must become separated from the top of the Mehrten by the introduction of a wedge of post-Mehrten, prepediment sediments. The rate of separation is probably slight. If it is the same as that on the lower Merced, 25 feet per mile, the top of the wedge goes below sea level 1.7 miles west of A. The sediments were presumably deposited above sea level, and, if so, the tilt axis passed some distance east of the wedge top's sea-level position. As this distance is unknown, it is assumed that A marks the axis.

The rate of uplift found for the G–H′ segment of the Early American channel in the following section is 37.3 feet per mile. If this is applied to E–F, the pretilt slope of that stretch is the incredibly low amount of 2.5 feet

[3] These logs, with top of Mehrten interpretations by Frank Olmsted, are on file in the Sacramento office of U. S. Geological Survey.

per mile. Therefore the tilt of 25.2 feet per mile determined for the pediment is assumed to be the proper factor to use as far east as F, a point chosen because it is on the contact between granite on the west and Mesozoic metamorphic rock on the east. The distance A–F on the projection being 12.4 miles, the uplift of F vs. A was (25.2)(12.4) = 312 feet.

F TO H: The tilting of this segment of the range was estimated by assuming that the original gradient of the G–H′ reach of the Early American was equal to that of the g–h′ stretch of the present river, namely (470 – 425)/(2.0) = 22.5 feet per mile. Using this factor and the 2.5-mile channel distance from G to H′, the pretilt difference in elevation between these points was (2.5)(22.5) = 56 feet. As the present rise from G to H′ is (1020 – 923) = 97 feet, the uplift of H′ vs. G caused by tilting was (97 – 56) = 41 feet; as the two points are 1.1 miles apart on the projection, the rate of tilting was (41)/(1.1) = 37.3 feet per mile. The distance from F to H measured on the S. 70° W. projection is 3.8 miles, and, accordingly, the uplift of H vs. F was (37.3)(3.8) = 142 feet.

H TO K: Point i on the modern channel is on the contact between Paleozoic and Jurassic rocks. The presence of a gold-quartz lode a short distance upstream shows that this contact was the locus of faulting in pre-Tertiary time. On the assumption that this was also a locality of yielding during the post-Early American deformation, the next turning point for the calculations is placed at I, 0.4 mile east of i and directly above the lode. J is chosen as a turning point because it lies at a marked change in slope on the ancient channel. It is assumed that bending occurred at I and J. K is selected as the terminal point of this set of calculations because it is immediately above a body of serpentine, along which movements that deformed the Pleistocene channel may have occurred.

The calculations at the top of the next column are based on the assumption that the pretilt gradient from H to K was the same as the evenly graded h–k stretch of the modern river, namely 20 feet per mile.

K TO L: The sharp nick in the ancient channel at L led to the choice of that place as a turning point in the calculations. If the tilt factor developed for J–K is applied to the K–L segment, the pretilt gradient of the latter reach will be found to have been 2.4 feet per mile, eastward. The two segments must have been tilted at different rates. The writer assumes that the tilting was accompanied by bending at a serpentine body, which is crossed by the channel a short way east of K. For the purposes of the calculation it is assumed that the bending took place at K.

	Distance, miles		Elevation, feet	
	On channel	On projection		Diff.:
H			1124	
	7.5	5.9		238
I			1362	
	4.5	2.6		458
J			1820	
	8.2	5.7		634
K			2454	

	Original rise, feet	Uplift, feet	Tilt, feet per mile
H–I	7.5 × 20 = 150	238 – 150 = 88	88/5.9 = 14.9
I–J	4.5 × 20 = 90	458 – 90 = 368	368/2.6 = 141.5
J–K	8.2 × 20 = 164	634 – 164 = 470	470/5.7 = 82.5

Subjacent to K–L is an evenly graded stretch of the modern stream, from k to the mouth of Rubicon River, with gradient (1150 – 925)/(6) = 37.5 feet per mile. Using this as the pretilt slope of K–L, the original rise along this 7.4-mile section was (7.4)(37.5) = 277 feet. The present rise from K to L is 2848 – 2454 = 394 feet, so that the uplift of L vs. K was 394 – 277 = 117 feet. As the distance L–K is 5 miles on the S. 70° W. projection, the rate of tilting of this segment of the Sierra was (117)/(5) = 23.4 feet per mile.

L TO N: At n the present river crosses a north-northeast-trending contact between granite on the east and metamorphic rock on the west, which is believed to have been the locus of folding or faulting of middle Eocene gravels (Hudson, 1955, p. 856). At least part of this deformation must have occurred after the cutting of the Early American channel. Accordingly, N, on the ancient channel and vertically above n, was chosen as a turning point in the calculations.

The profile of the L–Q stretch of the early stream is notably different from that of the subjacent part, l–q, of the modern river. The former is convex upward, whereas most of the latter is concave. Moreover, the over-all gradient of this section of the early channel is lower than that of the underlying stretch of the present stream. Accordingly, the profiles of the two channels contain no comparable reaches whose slopes might be the bases for computing uplift. The only recourse is to assume that the tilt factor of 23.4 feet per mile developed for K–L is properly applicable to

L–N. This is supported by the striking likeness of the L nick to that at l on the modern stream. Both appear to be purely erosional.

When the K–L tilt factor is applied to the 12.0-mile length of the L–N course on the S. 70° W. projection, the uplift of N with respect to L, caused by tilting, is (23.4)(12.0) = 281 feet.

N TO SUMMIT OF THE RANGE: Upstream from N, past R, the stream traverses a crustal block on which lie stations 3, 4, 5, and 6 of the early Tertiary Yuba River (Hudson, 1955, p. 861). The post-Middle Eocene tilt of this block was estimated to have been 130.4 feet per mile northwest, on a N. 35.9° E. axis, or in a N. 54.1° W. direction. Assuming internal rigidity, the uplifts of stations east of N since Middle Eocene were:

Station	Distance from N on N. 54.1° W. projection d	Uplift vs. N (d) (130.4)
Q	nil	nil
R	1.55 mi.	202 ft.
Granite Chief	4.59	599

The distances on the N. 54.1° W. projection used in the foregoing calculations cannot be read from Figure 5B, which is on a S. 70° W. projection throughout.

TABLE 2.—POST-EARLY AMERICAN UPLIFT

Stations	Distance on projection (miles)	Uplift		
		Rate (feet per mile)	Amount (feet)	
			Per course	Cumulative
	S. 70° W. projection			
A				..
	12.4	25.2	312	
F				312
	3.8	37.3	142	
H				454
	5.9	14.9	88	
I				542
	2.6	141.5	368	
J				910
	5.7	82.5	470	
K				1380
	5.0	23.4	117	
L				1497
	12.0	23.4	281	
N				1778
	N. 54.1° W. projection			
	1.5	60.0	90	
R				1868
	4.6	60.0	186	
Summit				2054

Only a part of the tilting of the Eocene deposits can be attributed to the post-Early American deformation because, in the Donner Pass region, a few miles north, several earlier episodes of deformation occurred at dates ranging from upper Eocene to middle Pliocene (Hudson, 1951, Table 1). The amount of this part of the tilting may be estimated by having recourse again to the gradient of a subjacent part of the modern stream as a basis for the calculation. Reaches Q–R and q–r (Fig. 5A) seem to be analogous. The over-all gradient of q–r is (5400 − 5050)/(7.2) = 48.6 feet per mile. If this was the Q–R gradient in Early American time, the rise from Q to R was then (48.6)(7.0) = 340 feet. The present rise is 6010 − 5580 = 430 feet, so that the increase in elevation of R vs. Q, caused by post-Early American tilting, was 430 − 340 = 90 feet. The distance Q–R on the N. 54.1° W. projection being 1.5 miles, the rate of tilt was (90)/(1.5) = 60 feet per mile. If this factor is applied to the stretch from N to the summit, which measures 4.6 miles on the N. 54.1° W. projection, the post-Early American uplift of Granite Chief vs. N is (60)(4.6) = 276 feet.

A TO GRANITE CHIEF: The calculations of uplift along the Early American channel are summarized in Table 2. Segments K–L and N–R of this stream overlie parts of the middle Eocene Yuba River. The amounts of post-Early American uplift calculated for K, L, N, R, and

TABLE 3.—COMPARISON OF POST-EARLY AMERICAN AND POST-MIDDLE EOCENE CALCULATED UPLIFTS

Stations		Uplift (feet)		
Early American	Corresponding points on middle Eocene Yuba	Post-Early American	Post-middle Eocene modified values	
			Average	Extreme additive
K	Station 10	1380	984	1245
L	Between stations 8 and 9	1497	1349	1590
N	Station 6	1778	1690	2146
R	4.3 mi., N. 47° E. from station 5	1868	1892	2348
Granite Chief		2054	2289	2745
..	Station 1	..	1887	2343
..	Donner Summit	..	1986	2442

Granite Chief are compared in Table 3 with the post-middle Eocene uplift of analogous points, derived from the study (Hudson, 1955) of the older channel's gradients. Since the and late Tertiary times, before the establishment of the Early American channel and an additional 2054 feet thereafter, making the total post-middle Eocene uplift 2745 feet.

TABLE 4.—BASAL GRAVELS IN CHANNEL OF MIDDLE EOCENE YUBA RIVER

Station	Locality	Distance from mouth (miles)	Nature of gravel
37	Sicard Flat	3	Spherical boulders of diabase, up to 2 feet in diameter, in quartzose grit
31	North San Juan	28	Largest boulder 4 x 1.3 x 1.3 feet (Gilbert, 1917, Pl. VIIA)
20	Thompson Hill	55	Cemented gravel averaging 8 inches in diameter, but with many 10-foot boulders* (Lindgren, 1911, p. 144)
19	Indiana Hill	58	Largest boulders have 1 foot mean diameter
9	Michigan Bluff	81	Quartz gravel with huge boulders at base* (Lindgren, 1911, p. 152)
1	Northwest of Donner Pass	109	Granite boulders up to 3 feet in diameter, in coarse grit
..	Mountain Gate	..	On tributary from north, joining main stream at station 9. Largest boulders 6 feet in diameter (Browne, 1890, p. 446)

* These exceptionally large boulders are at the mouths of steep tributaries and are not considered indicative of the competent slope of the main stream.

middle and lower courses of the two streams were widely divergent and the estimates of uplift along them were made by utterly different methods, the two sets of results appear to be mutually substantiating.

The writer has shown (1951; 1955) that the summit region was affected by several episodes of deformation in middle and late Tertiary times. Therefore the amount of post-Early American uplift should be considerably less than that since middle Eocene time. If one accepts this idea, the data of Table 3 indicate that the "average" modified values of post-middle Eocene uplift are not so near the truth as the "extreme additive" modified values. That the latter are the better choice is borne out by calculations based on two additional sets of physical data, set out in the following sections. Anticipating this confirmation, the writer concludes that the range summit at Granite Chief was raised 691 feet in middle

Gradients of the Middle Eocene Yuba River, Based on Size of Largest Boulder Transported

The cited amounts of post-middle Eocene uplift were found by setting up for each reach of the old channel an equation of two known quantities (reach direction and present gradient) and three unknown quantities (original gradient and rate and direction of tilt). The unknowns were found by the simultaneous solution of the equations of three consecutive reaches. The worth of the results of this method is supported by the following calculations of pretilt gradients, based on the size of the largest boulder found at various localities on the ancient channel.

The deposits in the gutter of the ancient Yuba, at seven places 3–109 miles above the mouth of that stream, are described in Table 4. The following estimate of the gradients required to move such detritus is based on the

assumption that the old river had the same tractive ability as the modern streams of the region.

TABLE 5.—LARGEST BOULDER TRANSPORTED ON VARIOUS SLOPES OF MODERN YUBA AND BEAR RIVERS

Locality	Upstream slope (feet per mile)	Largest boulder, mean diameter (feet)
1. Gravel pit on Yuba flood plain, 0.12 miles east of north end of "D" Street bridge, Marysville	>9.4	0.64
2. Yuba River at barrier, 14 miles east of Marysville	16.0	1.14
3. Yuba River at Parks Bar bridge, 1.5 miles above barrier	17.3	1.33
4. Bear River, just below bridge on Highway 49, 8.5 miles north of Auburn	33.0	3.00
5. Middle Fork of Yuba River, at Freeman's bridge (old site)	40.0	3.79

1. The 9.4 slope is that of 1909. For explanation of this choice, see text. Size: Graydon Oliver (Personal communication)

2. Size estimated from photograph (Gilbert, 1917, Pl. XXV B) of gravel under barrier. Slope: Gilbert (1917, p. 53)

3. and 4. Size from observation of boulders on bedrock of channels, October 1945 and April 1958

5. Size estimated from photograph of boulders on bedrock (Gilbert, 1917, Pl. IIB)

Gradients for 3, 4, and 5: U. S. Geological Survey quadrangle maps, Smartville (1951), Lake Combie (1950), and Camptonville (1948)

Table 5 contains the data of the largest boulder seen, and the upstream gradient, at four places on Yuba River and one site on Bear River, a parallel stream 10 miles to the south. The use of hundredths of a foot for the diameters in this table is not indicative of unnecessary precision in measurement but is the result, in part, of changing inches to feet and, in part, of calculating the mean diameters of ellipsoidal boulders.

The selection of competent gradient for the boulders found at locality 1 needs explanation. The gravel pit is situated on the flood plain of Yuba River, which is underlain by 19 feet of tailings from the hydraulic mines of the Sierra. The peak of the debris flood reached Marysville in 1905. Since then the river has cut a narrow channel 15 feet deep into the debris deposit but without notably affecting the flood plain. The flood of debris moved as a wave, the front

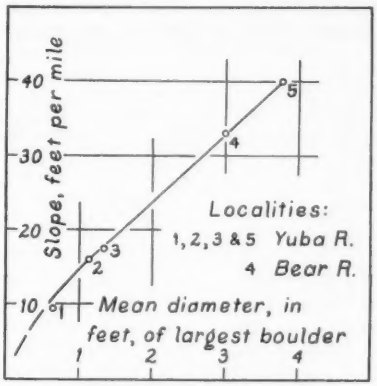

FIGURE 6.—RELATION OF BOULDER SIZE TO CHANNEL SLOPE, YUBA AND BEAR RIVERS

of which was inclined more steeply downstream than the rear. Boulders much larger than those that could move on the surface upstream from the crest were carried downstream on the front of the wave, and the competent slope for the transport of such boulders was that of the wave front. This was the 1905 river gradient at the site of the pit, an unknown amount, but believed to be slightly greater than the 1909 slope of 9.4 feet per mile.

The relation between the mean diameter of the largest boulder and the competent slope for its transport can be learned by plotting the data of Table 5 on square-grid graph paper (Fig. 6). The plotted points for localities 3 and 4 fall almost exactly on the straight line connecting the points determined from the data of localities 2 and 5, so that this line is a sound representation of the relation of maximum boulder diameter to competent gradient for diameters as large as, and larger than, the 1.14 feet of locality 2. For smaller diameters the graph must bend downward, in order to reach the origin. It was drawn to pass a little above the point plotted for locality 1, to conform with the idea that the 1905 pertinent slope at that place was slightly greater than the known 1909 gradient of 9.4 feet per mile. The Figure 6 graph is based, not on selected data, but on all available information.

Gradients estimated from the Eocene gravel data of Table 4, by means of this graph, are shown in Table 6, with the slopes at the same localities found by the triplet equation method

TABLE 6.—COMPARISON OF GRADIENTS ON MIDDLE EOCENE YUBA RIVER, CALCULATED BY BOULDER-DIAMETER AND TRIPLET-EQUATION METHODS

Station		Diameter of largest boulder (feet)	Gradients (feet per mile)		
			By boulder diameter method (Fig. 7)	By triplet method (Hudson, 1955, Table 7) Modified	
				Average	Extreme
37	Sicard Flat	2.0	23.8	29.2	27.1
31	N. San Juan	1.9	22.9	34.9	29.8
19	Indiana Hill	1.0	14.8	17.0	11.0
1	Donner Pass	3.0	33.0	212.5	212.5
..	Mountain Gate	6.0	60.0	63.5*	..

* Based on assumption that this channel underwent same tilt as that calculated for triplet 8-9-10-11

(Hudson, 1955, Table 7). Since the largest boulder at any locality probably would not be seen, the boulder-size method must, in most cases, give results that are less than the true gradients. Thus the solutions for four of the localities indicate that the results of the boulder-size method support those determined by triplet equations and that the "extreme additive" modified values of post-middle Eocene uplift are nearer the truth than those which were called, on mathematical grounds, the "average" modified values. The extreme disparity between the results of the two methods at station 1 is thought to be of little moment, because: (1) The 3-foot boulders observed here may not represent the maximum size, (2) these boulders may represent what were delivered to station 1 on a gradient much gentler than that below that point, and (3) the Figure 6 graph may not be useful for the headwaters of a Sierra river.

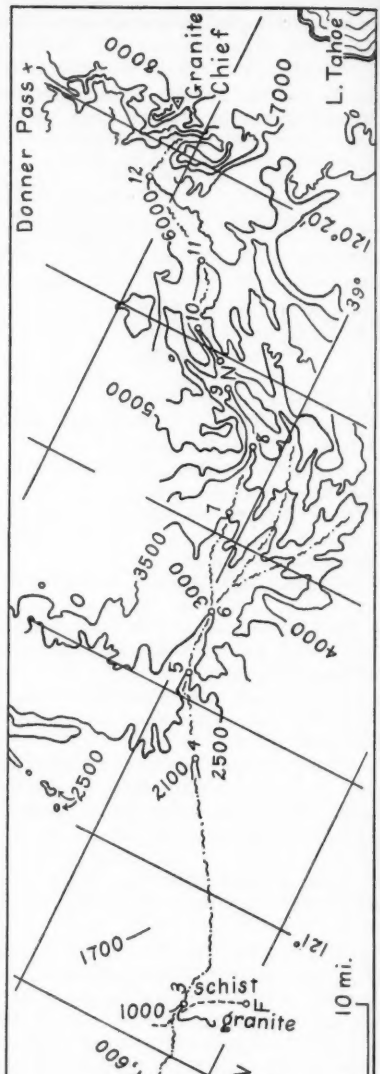

FIGURE 7.—LATE MIOCENE (PRE-MEHRTEN) TOPOGRAPHY, WEST OF GRANITE CHIEF

Post-Mehrten Uplift

The Mehrten andesitic deposits buried a late Miocene landscape which was underlain by pre-Cretaceous crystalline rocks, middle Eocene gravels, and acidic tuffs of Oligocene or early Miocene age. The shape of this land surface, in the region southwest of Granite Chief, was found by drawing structural contours of the

bottom of the Mehrten, based on the geologic maps of the Placerville quadrangle (Lindgren and Turner, 1893) and the Truckee and Colfax quadrangles (Lindgren, 1897; 1900). A late Miocene stream flowed southwest from the vicinity of Granite Chief (Fig. 7). The data of this channel are given in Table 7 and Figure 8 and are shown in N. 70° E. projection by the A′–2–12 graph in Figure 5B.

TABLE 7.—LATE MIOCENE CHANNEL

Stations	Present elevation (feet)	Course	Channel		Distances on projections	
			(miles)	(feet per mile)	S.70°W. (miles)	N.54°W. (miles)
Granite Chief					2.8	1.8
13	6500	N. 81° W.	1.8	278	1.6	1.6
12	6000	S. 31° W.	5.6	89	4.3	−0.5
11	5500	S. 66° W.	4.1	122	3.8	1.9
10	5000	S. 33° W.	2.1	129	1.7	−0.2
N	4730	S. 43° W.	1.6	144	1.5	4.6
9	4500	S. 41° W.	3.6	139	3.0	
8	4000	S. 83° W.	4.0	125	3.9	
7	3500	S. 75° W.	5.8	86	5.5	
6	3000	S. 84° W.	3.8	132	3.6	
5	2500	S. 57° W.	4.9	82	4.8	
4	2100	S. 68° W.	13.8	80	13.7	
3	1000	S. 80° W.	5.2	77	36.0	
2Y	600	S. 46° W.	1.25	80	5.0	
2	500	S. 6° W.	1.6	62	1.2	
2X	400	S. 21° W.	3.4	69	0.6	
1	200	S. 17° W.	4.1	49	2.1	
A′	sea level				2.5	
					11.4	

These data are far less useful for uplift calculations than those of the middle Eocene Yuba and Early American rivers. Not one set of [illegible] reaches will yield a solution 1569, column 2, line 12 from bottom, stream, and the lack of character in the late Miocene profile precludes the reach by reach comparison with the modern Middle Fork that was found useful in the calculation of post-Early American uplift. Despite these shortcomings the writer believes that the results of the following calculations are credible.

The basic assumptions are: (1) The post-Mehrten tilting took place on the same axes as those selected for the post-Early American movements—*e.g.*, a N. 20° W. axis for the A′–N block and a N. 36° E. axis for the block from N to Granite Chief; (2) Deformation of the late Miocene channel occurred at places corresponding to F and N of the Early American channel, namely, at 3 and N; (3) Despite the likelihood that the late Miocene channel was deformed at several places between 3 and N, the tilt calculation may be based on the over-all rise from 3 to N.

A′ TO 3: In calculating the post-middle Eocene tilting along the Tertiary Yuba River, the writer showed (1955, p. 844) that

$$S: - S_0 = T \cdot \sin_1 \delta_1,$$

equation should read $S_1 - S_0 = T \cdot \sin \delta_1$ [illegible] gradients of a reach of a [illegible] te of tilting, and δ_1 is the angle between the bearing of the reach and that of the tilt axis. For the reaches 2X–2 and 2–2Y of the pre-Mehrten stream,

$$62 - S_0 = T \cdot \sin (20° + 6°) = T \cdot \sin 26° = 0.438\,T$$
$$80 - S_0 = T \cdot \sin (20° + 46°)$$
$$= T \cdot \sin 66° = 0.914\,T.$$

Assuming that the two reaches had a common S_0 and a common T, S_0 may be eliminated by

FIGURE 7.—LATE MIOCENE (PRE-MEHRTEN) TOPOGRAPHY, WEST OF GRANITE CHIEF

both
ma
and
qua

PLATE 7. ZION'S CANYON (LOOKING EAST) TOPOGRAPHY. WEST OF GRANITE CLIFF

bottom of the Mehrten, based on the geologic maps of the Placerville quadrangle (Lindgren and Turner, 1893) and the Truckee and Colfax quadrangles (Lindgren, 1897; 1900). A late Miocene stream flowed southwest from the vicinity of Granite Chief (Fig. 7). The data of this channel are given in Table 7 and Figure 8 and are shown in N. 70° E. projection by the A′–2–12 graph in Figure 5B.

TABLE 7.—LATE MIOCENE CHANNEL

Stations	Present elevation (feet)	Course	Channel		Distances on projections	
			(miles)	(feet per mile)	S.70°W. (miles)	N.54°W. (miles)
Granite Chief					2.8	1.8
13	6500	N. 81° W.	1.8	278	1.6	1.6
12	6000	S. 31° W.	5.6	89	4.3	−0.5
11	5500	S. 66° W.	4.1	122	3.8	1.9
10	5000	S. 33° W.	2.1	129	1.7	−0.2
N	4730	S. 43° W.	1.6	144	1.5	4.6
9	4500	S. 41° W.	3.6	139	3.0	
8	4000	S. 83° W.	4.0	125	3.9	
7	3500	S. 75° W.	5.8	86	5.5	
6	3000	S. 84° W.	3.8	132	3.6	
5	2500	S. 57° W.	4.9	82	4.8	
4	2100	S. 68° W.	13.8	80	13.7	
3	1000	S. 80° W.	5.2	77	36.0	
2Y	600	S. 46° W.	1.25	80	5.0	
2	500	S. 6° W.	1.6	62	1.2	
2X	400	S. 21° W.	3.4	69	0.6	
1	200	S. 17° W.	4.1	49	2.1	
A′	sea level				2.5	
					11.4	

Miocene profile precludes the reach by reach comparison with the modern Middle Fork that was found useful in the calculation of post-Early American uplift. Despite these shortcomings the writer believes that the results of the following calculations are credible.

The basic assumptions are: (1) The post-Mehrten tilting took place on the same axes as those selected for the post-Early American movements—*e.g.*, a N. 20° W. axis for the A′–N block and a N. 36° E. axis for the block from N to Granite Chief; (2) Deformation of the late Miocene channel occurred at places corresponding to F and N of the Early American channel, namely, at 3 and N; (3) Despite the likelihood that the late Miocene channel was deformed at several places between 3 and N, the tilt calculation may be based on the over-all rise from 3 to N.

A′ TO 3: In calculating the post-middle Eocene tilting along the Tertiary Yuba River, the writer showed (1955, p. 844) that

$$S_1 - S_0 = T \cdot \sin_1 \delta_1,$$

in which S_1 and S_0 are the present and pretilt gradients of a reach of a channel, T is the rate of tilting, and δ_1 is the angle between the bearing of the reach and that of the tilt axis. For the reaches 2X–2 and 2–2Y of the pre-Mehrten stream,

$$62 - S_0 = T \cdot \sin (20° + 6°) = T \cdot \sin 26° = 0.438\,T$$
$$80 - S_0 = T \cdot \sin (20° + 46°)$$
$$= T \cdot \sin 66° = 0.914\,T.$$

Assuming that the two reaches had a common S_0 and a common T, S_0 may be eliminated by

subtracting one equation from the other to form

$18 = 0.476\ T$, or $T = 37.8$ *feet per mile*

and $S_0 = 80 - (0.914)(37.8)$

$= 80 - 34.5 = 45.5$ *feet per mile.*

Yuba River has a like position. The writer assumed that the 10–11 and 11–12 courses had a common original slope and that they were tilted on the same N.36°E. axis as that found

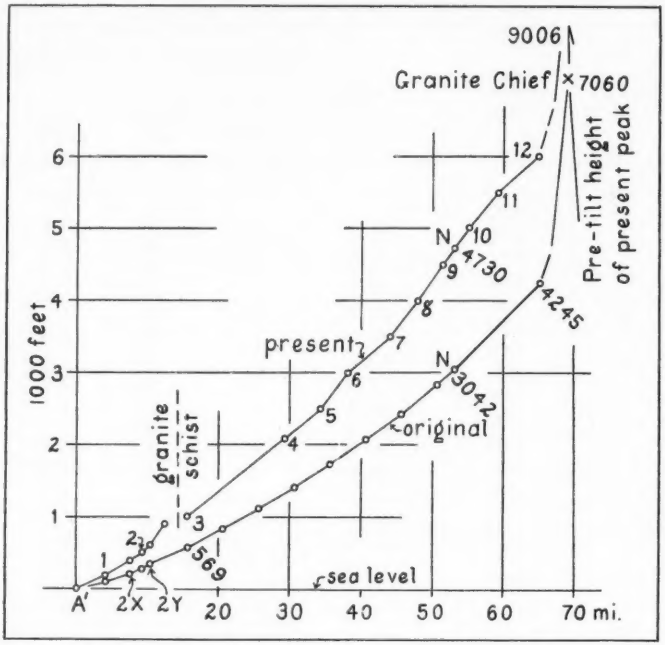

FIGURE 8.—LONG PROFILES OF LATE MIOCENE STREAM, WEST OF GRANITE CHIEF

Assuming, as in the post-Early American uplift calculation, that the tilting was S. 70° W. on a sea-level axis, the writer found the position of the axis (Fig. 7) by drawing a line N. 20° W. from a point between two closely adjacent wells in which the base of the Mehrten was determined at −6 and +13 feet with respect to sea level. The location of point A′, the intersection of the channel with the axis, may be much in error, but this does not affect the calculation, which depends, not on the position of the point, but on that of the axis, which lies 11.4 miles from station 3 along the S. 70° W. projection. If the tilt factor, $T = 37.8$, is applied to this distance, the uplift of 3 with respect to A′ is (11.4)(37.8) = 431 feet, making the pretilt elevation of station 3 1000 − 431 = 569 feet.

N TO GRANITE CHIEF: The pair of late Miocene reaches, 10–11 and 11–12, lies on the western part of this block, and the southern end of the 3–4–5–6 triplet of the middle Eocene for the Eocene triplet (Hudson, 1955, p. 861). The 10–11 reach runs S.66°W. with present gradient 122 feet per mile, and the 11–12 course runs S.31°W. on a slope of 89 feet per mile. The following calculation is similar to that of the preceding section:

$122 - S_0 = T \cdot \sin(66° - 36°) = T \cdot \sin 30° = 0.5\ T$

$89 - S_0 = T \cdot \sin(31° - 36°)$

$= T \cdot \sin(-5°) = -0.087\ T$

$33 = 0.587\ T$, $T = 56.2$ *feet per mile*

$S_0 = 122 - (0.5)(56.2)$

$= 122 - 28 = 94$ *feet per mile.*

The distance from N to Granite Chief being 4.6 miles on the N.54°W. projection, the uplift of the peak with respect to N was (56.2)(4.6) = 258 feet.

3 TO N: To learn the pretilt shape of the long profile of this part of the late Miocene stream, the writer assumed that, in the 51-mile distance between the mid-points of courses 2X–2–2Y

and 10–12, the gradient increased, as a simple function of distance, from the 45.5 feet per mile found for the former course to the 94 feet per mile determined for the latter. Equations set up by plotting these data of slope and distance as

TABLE 8.—POST-MEHRTEN UPLIFT COMPARED WITH OTHERS
(Distances in feet)

Stations	Post-Early American	Post-Mehrten	Post-middle Eocene modified values	
			Average	Extreme
F and 3.....	312	431	..	..
N.....	1778	1688	1690	2146
Granite Chief...	2054	1946	2289	2745

straight lines on square grid and semilogarithmic graph charts yielded 2602 and 2473 feet, respectively, for the pretilt rise from 3 to N. The latter value is adopted because it leads to a maximum amount of uplift. As the present rise from 3 to N is 3730 feet, the uplift of N vs. 3 was 3730 − 2473 = 1257 feet.

A′ TO GRANITE CHIEF: The results of the foregoing calculations are summarized graphically in Figure 8 and are shown in Table 8 compared with the values found for the post-middle Eocene and post-Early American uplifts. The latter values are well supported by the 1946-foot post-Mehrten uplift determined for Granite Chief, and by the results for intermediate stations between A′ and the summit.

Uplift Determined by Paleobotanical Means

Fossil plant remains from three localities within 50 feet of the base of the Mehrten andesitic deposits in the Carson Pass area were determined by Axelrod (1957) to be of Mio-Pliocene age. In this region on the summit of the Sierra, 39 miles south-southeast from Granite Chief, the base of the Mehrten is 9000 feet above sea level. By comparison with living floras of the western slope of the range, he decided that the Carson Pass localities were about 2500 feet above sea level when Mehrten deposition began and concluded (his Table 2) that the basement-volcanic contact has been uplifted 9000 − 2500 = 6500 feet since that time. These are the only known Tertiary fossil localities in the higher parts of the Sierra. In the absence of direct evidence he inferred that, before Mehrten deposition, the crest of the range declined north of Carson Pass, in a distance of 47 miles, to an elevation of about 2000 feet at Donner Pass. This was based on the idea that, had the range been higher, there would have been such a rain shadow east of the summit that the rainfall in west-central Nevada would have been less than that needed to support the Mio-Pliocene flora known to have grown there.

As the base of the Mehrten at Donner Pass is now at 7300 feet elevation, Axelrod concluded that the post-Mehrten uplift there has been 7300 − 2000 = 5300 feet. The distances from Carson Pass to Granite Chief and Donner Pass being, respectively, 39 and 47 miles, this seems to call for a post-Mehrten uplift at Granite Chief of 6500 − (6500 − 5300) (39/47) = 5505 feet. This is from 2 to nearly 3 times the results yielded by the physical methods for post-Early American, post-Mehrten, and post-Middle Eocene uplift.

This wide disparity appears to preclude compromise. The writer is not equipped to criticize Axelrod's work on either botanical or climatological grounds. But the facts that the results obtained from three dissimilar physical methods, based on three independent sets of data, are mutually supporting, and that one of these is partially substantiated by a fourth method, are thought to show that the amounts of uplift estimated by these means deserve credence. Therefore the paleobotanical work should be critically reviewed.

The Mio-Pliocene fossil plant localities of west-central Nevada, listed by Axelrod (1957, Fig. 1), are from 70 to 120 miles southeast, east, and northeast of Donner Pass, at altitudes of about 4000 feet. Physiographic and other evidence suggests that the lower parts of this region have been undergoing absolute depression for some time and the higher parts have been relatively stable. A striking example is the change wrought on Stingaree Valley, 120 miles east of Donner Pass, at the time of the Fairview Peak earthquake in 1954 (Slemmons, 1957). The scarps that formed on either side of this 4-mile-wide valley indicate that it is an active graben between the Clan Alpine range on the east and low hills on the west. More important, a redetermination of elevations along the course of Highway 50, across the valley, showed that, in absolute measure, the graben dropped a maximum of 5.2 feet, while the block to the east stood still and that to the west rose only 0.2 foot. For further evidence on similar changes in elevation in the western Great

Basin, since the end of Tertiary time, *see* Lindgren (1911, p. 41), Lawson (1912, p. 195), Matthes (1930, p. 30), and Hudson (1955, p. 859).

On the basis of the foregoing, one might conclude that the original altitudes of the sites of these Mio-Pliocene floras were more than 4000 feet, instead of 2000–2500 feet, as maintained by Axelrod (1957, p. 33). But nothing is known counter to the idea that these localities were raised about 2000 feet before the beginning of the period of sinking, perhaps in the manner indicated in Le Conte's theory (1889, p. 262)—*i.e.*, at the end of the Tertiary, what is now the Great Basin was raised as a huge arch which concurrently collapsed into a series of basins and ranges. Physical proof of the 2000-foot uplift would substantiate the idea of the low environment of these floras, but unless such proof is offered the writer believes that Axelrod's estimate of the uplift at Donner Pass cannot withstand the impressive array of adverse physical evidence.

Nature of Internal Deformation of the Range

The writer assumed in the calculations that the post-broad-valley and post-Early American uplifts were accompanied by bending at certain localities where pre-Tertiary rocks are faulted. At some of these places possibly the Pleistocene channels have been faulted rather than bent. The most notable of these is between I′ and I″ (Fig. 5A) on the Early American channel, where the hanging-valley evidence suggests that I″ has been raised 135 feet with respect to I′ in a distance of 0.3 mile. But it seems unlikely that faulting affected the Pleistocene channel here, or at other less noteworthy localities, since it would have produced scarps, which are not now to be seen, and the amount of erosion needed to erase these scarps would be enough to destroy the hanging-valley evidence.

Summary of Conclusions

Using the hanging-tributary-valley method of Matthes (1930), the long profiles of his broad-valley and mountain-valley stages of the preglacial, Pleistocene Merced River can be extended downstream, from the positions determined by him in the Yosemite Valley region, to the western edge of the Sierra.

Similar channels, determined by the same method on the Middle Fork and main American River, are here called Early American and Later American stages.

The ancient Merced and American streams established pediments below their canyon mouths. The older of these on the Merced, China Hat pediment, is correlated with the top of Davis and Hall's (1959) Turlock Lake formation which, in their chronology, makes the pediment and the broad-valley stream that formed it, of late early Pleistocene age. The Early American channel is assigned to the same age because of its similarity to the broad-valley Merced, both in form of long profile and relation to the subjacent modern stream. The mountain-valley channel of the Merced is correlated with the unconformity at the top of Davis and Hall's Riverbank formation, thought by them to mark the top of the middle Pleistocene.

Since the establishment of the earlier channels the Sierra Nevada block was tilted west-southwest with accompanying internal deformation by bending.

Matthes' estimate of 9000 feet post-broad-valley uplift of the Sierra at the head of Merced River is unsound. A new calculation, based on the gradients of the modern stream and on the deformation of the Pleistocene sediments at the mouth of the ancient river, gives 3930 feet for this uplift.

The uplift at Granite Chief at the head of Middle Fork of American River is estimated by similar means to have been 2054 feet since Early American (early Pleistocene) time. This is substantiated by calculations by wholly different methods, which indicate the uplift at the same locality to have been 2745 feet since middle Eocene time and 1946 feet since late Miocene.

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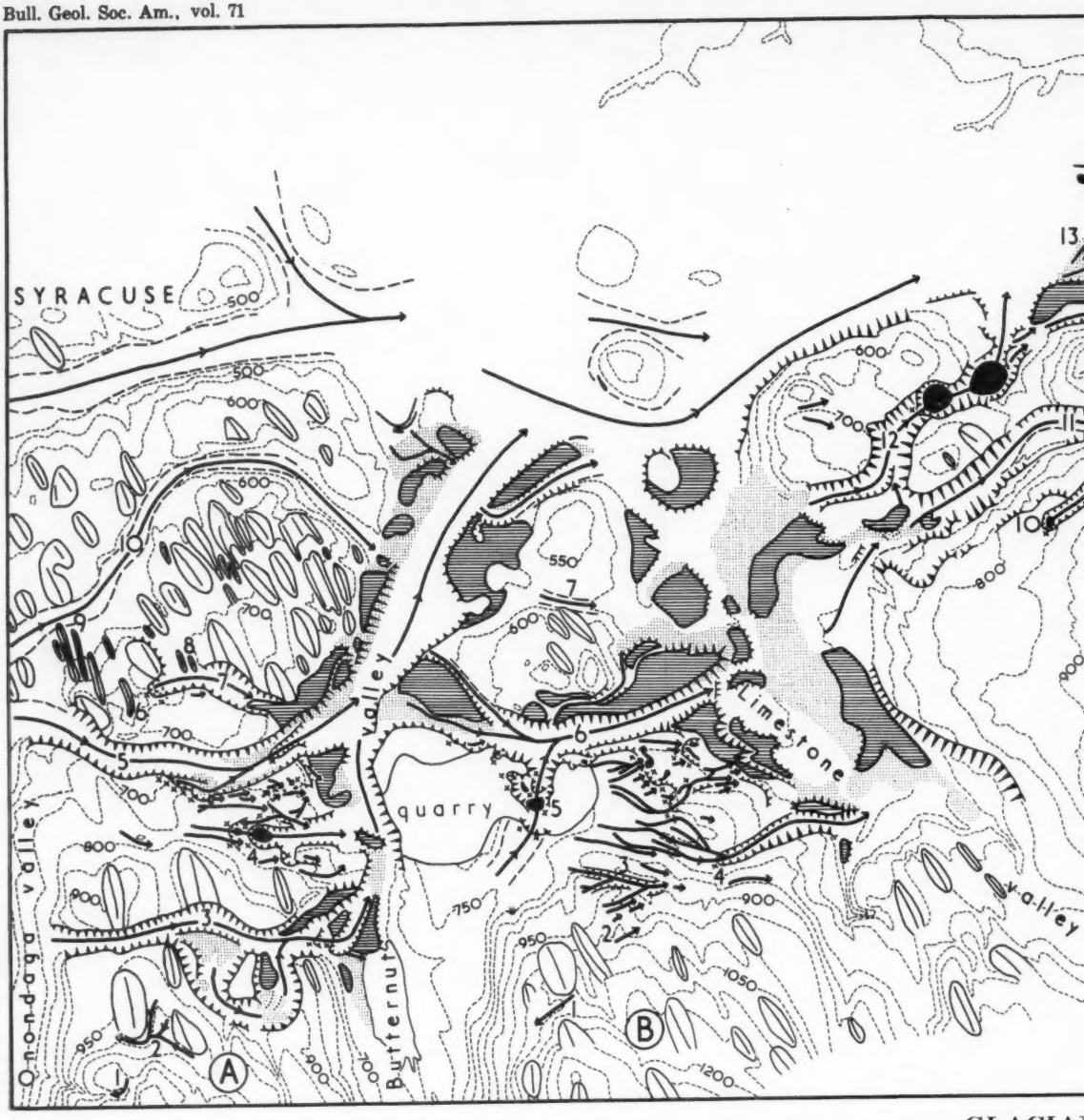


GLACIA

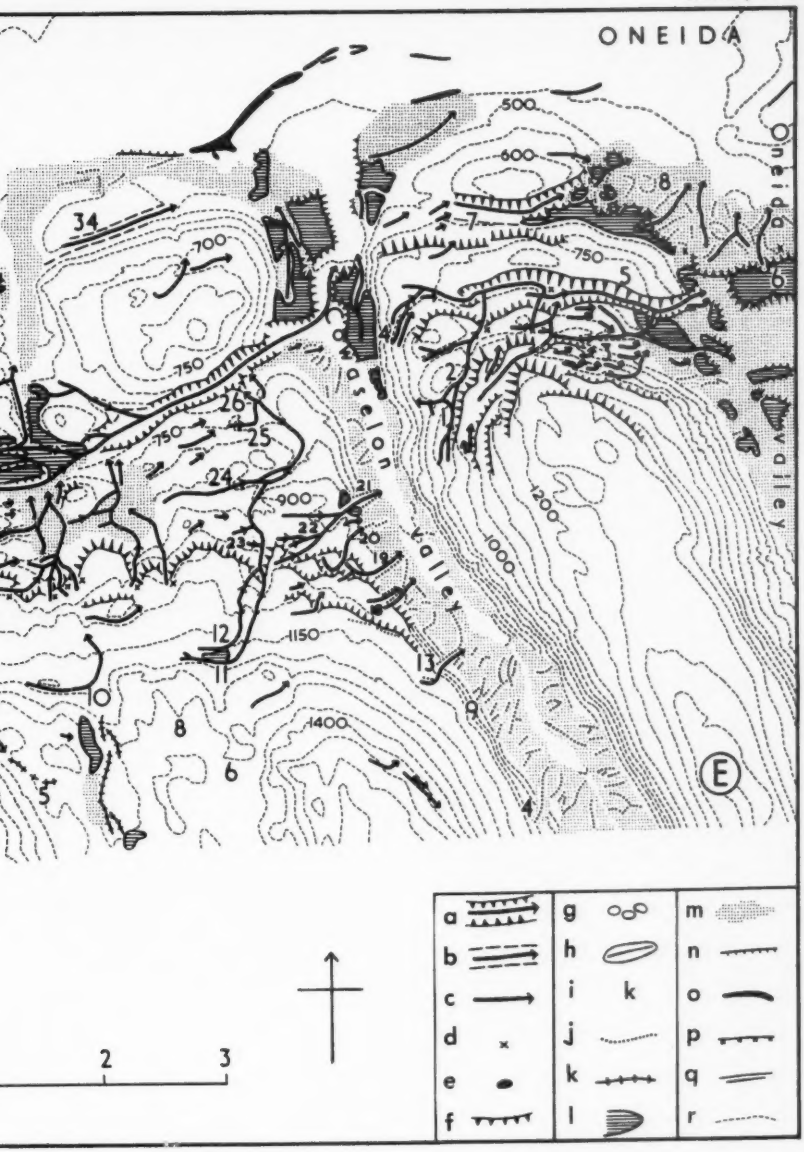
a, large steep-sided glacial-d
channel bounded by gentle to m
of flattish floor); c, other glacial-d
basin; f, steep slope of side of la
scarp, and bounding terrace; g,
ridge; k, esker; l, kame terrace,

CIAL LAND FORMS OF THE SYRACUSE-ONEIDA AREA, NEW YORK

cial-drainage channel; b, large glacial-drainage to moderate slopes (broken lines indicate limits cial-drainage channel; d, waterfall site; e, plunge of large glacial-drainage channel, of limestone e; g, end moraine; h, drumlin; i, kettle; j, kame ace, river terrace, outwash terrace, or delta surface; m, fluvioglacial sand and gravel (symbol omitted on terrace and delta surfaces); n, Lake Iroquois cliff; o, bar of Lake Iroquois; p, cliff line above Iroquois level; q, small shore features, mostly above Iroquois level; r contour, interval 50 feet (omitted where other symbols shown). Contours from U. S. Geological Survey maps.

RACUSE-ONEIDA AREA, NEW YORK

face; m, fluvioglacial sand and gravel (symbol omitted on terrace and delta surfaces); n, Lake Iroquois cliff; o, bar of Lake Iroquois; p, cliff line above Iroquois level; q, small shore features, mostly above Iroquois level; r contour, interval 50 feet (omitted where other symbols shown). Contours from U. S. Geological Survey maps.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SUBGLACIAL, MARGINAL, AND OTHER GLACIAL DRAINAGE IN THE SYRACUSE-ONEIDA AREA, NEW YORK

BY J. B. SISSONS

ABSTRACT

In his numerous papers on central New York, H. L. Fairchild interpreted almost all the glacial-drainage features as of marginal origin. In the Syracuse-Oneida area, however, there is morphological evidence of subglacial stream erosion by both local and Great Lakes waters, of subglacial deposition by local waters, and of englacial flow in places. Evidence also indicates that the Great Lakes waters in some places flowed in channels walled by ice. Only limited evidence of ice-dammed lakes exists in the Syracuse-Oneida area itself, except at the lowest altitudes, where numerous small shore features indicate a former local lake above the highest Iroquois shore line. There is evidence of two periods of Great Lakes drainage across the hillslopes of the Syracuse-Oneida area during the oscillatory retreat from the Wisconsin maximum. The writer suggests, however, that these two periods of Great Lakes drainage alone are insufficient to account for the vast, complex channel system.

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INTRODUCTION

"The glacier acting as a barrier to northward drainage is the fundamental fact to be apprehended by the reader. The ice sheet was a melting dam during both its advance and its retreat, and waters were flowing copiously from it, not into it. Valleys or land depressions sloping toward the ice front were by the ice barrier made into lake basins."

With these words Fairchild (1909) opened a paper on glacial drainage in central New York in which he summarized his conclusions given in many of his numerous earlier publications. The quotation shows that Fairchild, like many other writers in North America and Britain, erected his interpretation of glacial drainage on the assumption that glacier ice is an impenetrable barrier to waters outside it. This paper's main purpose is to show that during the last retreat of the ice from the Syracuse-Oneida area, meltwaters frequently flowed into, beneath, and across the ice, giving rise to a variety of erosional and depositional features. This recognition of the complexities of the glacial drainage necessitates certain modifications of Fairchild's interpretation of the glacial retreat, an interpretation that appears to have remained essentially unchallenged.

ACKNOWLEDGMENTS

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METHODS

The geomorphic features of the Syracuse-Oneida area were initially mapped from air

photographs, and this mapping was subsequently checked, and modified where necessary, by systematic traversing on foot. The generalized results of the mapping are shown on Plate 1, numerous details and minor features having been omitted in order to produce a map of reasonable dimensions.

Major Relief

The area lies on the northern margin of the Alleghany Plateau province. In the south the plateau is 1200 to 1700 feet in altitude, but the ground descends abruptly northward to the Iroquois Lake plain at about 400 feet above sea level. The upland is broken into five major sections by the deep, steep-sided valleys of the northward-flowing Onondaga, Butternut, Limestone, Chittenango, Cowaselon, and Oneida creeks. These five sections are lettered A to E in Plate 1, and points referred to in the text are numbered, *e.g.*, A3, E5. Since constant reference to this plate is required throughout the text, specific mention of it is generally omitted.

The bedrocks of the area are mainly limestones and shales dipping gently southward and southwestward. Shales crop out on the higher parts of the spurs, and in the southwestern part of the area their outcrop is roughly co-extensive with the drumlin region. Massive limestones beneath the shales correspond with a well-marked bench that rises eastward from an altitude of about 700–800 feet near Syracuse to about 1200–1300 feet near Oneida. Northward the limestone bench in many places terminates in a very steep escarpment, or in several escarpments, overlooking the lower ground to the north developed on the underlying shales.

Previous Work

The only comprehensive treatment of the glacial drainage of the Syracuse-Oneida area (and adjacent areas) is by Fairchild, who published numerous papers during a period of 35 years on aspects of the glaciation of central New York. His conclusions regarding the glacial history changed slightly as he accumulated evidence, but his final conclusions (1909; modified in 1932a; 1932b), which Flint (1953; 1956; 1957, p. 343–348) apparently accepted, were, briefly, as follows.

(1) As the ice sheet withdrew from the northern slopes of the uplands of central New York, lakes were impounded in the valleys of the Finger Lakes region and in the valleys to the east. The earliest drainage of these lakes was southward through the cols at the valley heads.

(2) This southward drainage was gradually replaced by westward drainage into Lake Warren, the outlet of which was the Grand River valley in western Michigan. The westward drainage certainly commenced as far east as the Onondaga valley and probably as far east as the Limestone valley.

(3) The westward drainage was replaced by eastward drainage, the latter developing first in the eastern part of central New York and slowly extending farther west. This eastward drainage did not involve the Warren waters for a considerable time, however, since the ice margin in the Batavia area (100 miles west of Syracuse) remained stationary and continued to hold back the Warren waters while it was retreating farther east. Consequently during this retreat Great Lakes waters did not escape across the hillslopes of the Syracuse-Oneida area, which were affected only by local waters.

(4) Eventually the ice margin in the Batavia area began to retreat, and the Warren waters were thus able to escape eastward, a series of channels being cut in the Batavia-Rochester area from 870 feet down to about 480 feet. These waters flowed only through the low ground in the Rochester-Syracuse-Oneida-Rome district, escaping through the Mohawk Valley.

(5) After retreating an unknown distance, the ice readvanced and Lake Warren was re-formed with its outlet again through the Grand River valley. This time, however, the ice margin was highest in the Syracuse area, not the Batavia area. Consequently, during this readvance and the final retreat of the ice, huge channels were cut near Syracuse by the large volumes of water escaping eastward to the Mohawk Valley.

(6) With continued retreat of the ice, Lake Iroquois was established at various altitudes up to 460 feet at Rome and 435 feet at Rochester (this difference in altitude being partly due to isostatic recovery of the area).

Local Glacial Drainage

Except in the extreme eastern part of the area, the features produced by locally derived waters occur mainly on the spurs and are found above a zone descending from about 900–950 feet in the west to about 650 feet in the east. They comprise numerous meltwater

channels and kames, a few small deltas, and two small esker systems.

Most of the channels are small, ranging from 2 or 3 feet deep to 40 or 50 feet deep. A few channels attain depths of 100 feet for short distances, but they are much narrower than the channels formed by Great Lakes waters. The channels were excavated by waters that ultimately found their way eastward to the Mohawk Valley, excepting a few in the extreme southwest that carried meltwaters westward. These channels (A1, A2, B1) are consistent with Fairchild's conclusion that drainage was at first westward on spurs A and B.

Although numerous marginal channels occur above the levels affected by Great Lakes waters it has not been possible to correlate them throughout the area studied and thus map successive positions of the ice margin. In fact, it is rarely possible to deduce the position of the ice margin from marginal channels for more than a mile in any one locality. The principal exception occurs on spur D and even there the ice-margin position can be traced only 3 miles. Channel D1 starts at about 1230 feet and ends in a delta whose surface lies at about 1220 feet. The shallow channel D2 begins at the edge of the delta and descends to about 1210 feet. A short distance to the east channel D3 begins at 1210 feet and ends in a delta whose surface lies between 1200 and 1170 feet. A mile farther northeast channel D7 commences at 1170 feet and terminates at about 1140 feet. This sequence thus provides an example of very small ice-dammed lakes draining from one to another by channels cut across spurs. Fairchild (1909, p. 39) described part of this sequence. Significantly, it appears to be unique above the levels of the region affected by Great Lakes waters in the area studied. According to the marginal-drainage hypothesis, one would expect innumerable examples, for at every position of the ice margin during its retreat, there should have been along or near the ice edge either meltwater erosion, producing channels, or deposition, forming kames, kame terraces, deltas, or lake-floor deposits. Fairchild recognized this, and his maps of the Syracuse-Oneida area (and of adjacent areas) show such features in abundance (Fairchild, 1909): in particular, numerous channels clearly convey the impression of meltwaters sweeping along the ice margin where it lay against the slopes of the successive spurs. Fairchild's maps are misleading in this respect, however, since many of the smaller channels marked on them do not appear to exist.

Plate 1 shows that there is no evidence of marginal drainage in numerous places where one would expect to find it on the basis of Fairchild's hypothesis. The absence of such evidence is by no means uncommon in similar localities elsewhere in North America and in Europe and has been frequently explained as due to (1) subsequent destruction of the evidence or (2) adjacent ice-dammed lakes having become confluent across intervening spurs. Regarding the first point, it is probable that a few small channels have been destroyed or obscured, but this explanation is clearly inadequate in many instances. For example, on the basis of the marginal-drainage hypothesis, one would expect series of marginal channels and related deposits to occur on spur D immediately below the level of the well-defined series D1, D2, D3, and D7 described heretofore. Yet only two shallow marginal channels and one small deltaic area occur here, the rest of this area apparently being devoid of marginal features. The second explanation might account for the absence of marginal features at a few points but can hardly be invoked repeatedly. Furthermore, it presupposes the existence of marginal lakes of considerable depth. There is very little evidence of even small marginal lakes having existed on the spurs, however, and no evidence at all of large ones: kames, deltas, and lacustrine deposits are of limited extent on the spurs.

If then, the absence of marginal features from many parts of the spurs cannot be explained in either of these ways, it would seem that one must conclude that the meltwaters frequently flowed beneath, through, and/or over the ice itself. Concerning the two last, evidence is obviously difficult to obtain. There is abundant evidence of the first, however. A considerable number of channels run along the contours of hillsides for distances ranging between a few hundred feet and about a mile before turning through a right angle, in places abruptly, to run straight down the slopes. Examples include C2, C4, D10, D11, D12, D16, and several channels in area 3 on spur E. The parts of such channels running along the slopes were cut by meltwaters flowing along the ice margin, and the down-slope segments were produced by these same waters flowing beneath the ice. The sharp bend in a channel of this type is the point at which the meltwaters turned to flow beneath the ice. Many such channels have been

described in northern Sweden by Mannerfelt (1945), who called them "subglacial chutes".

Some subglacial chutes carried beneath the ice the meltwaters from a number of marginal channels at different stages of the retreat. Channel system D14 is an example. The upper part of this system, aligned approximately west-east and cut about 10–20 feet into limestone, was formed along the ice margin. Traced eastward the main channel is found to increase abruptly in depth to about 80 feet as it swings around through 90° to run steeply down-slope. This down-slope segment, in part of subglacial origin, runs at right angles to several marginal channels, some of which are clearly developed on its western side but do not exist on its eastern side. Thus the down-slope channel must have carried the waters of these marginal channels beneath the ice. Other examples of channel systems of this type are shown in plate 1, especially on spur D.

This interpretation of certain valleys running down hillsides, including some valleys containing streams, as wholly or partly of subglacial origin is supported by other evidence. In the case of marginal channels that turn through a right angle to run down slopes, any other origin than a subglacial one for the down-slope section is difficult to visualize, for both parts of such channels are continuous with each other and form a single feature that can be explained only in terms of erosion by waters flowing from one end to the other (*i.e.*, marginally to the ice, then subglacially). Furthermore, the down-slope section of some of these channels ends abruptly part way down a steep slope (*e.g.*, around E3). It is difficult to see how a normal stream could suddenly cease to erode part way down a steep slope. On the other hand, the sudden cessation of bedrock erosion by a subglacial stream could be due either to its having become englacial or to its having descended to the level of the water table in the ice.

The presence of streams in some subglacial channels is to be expected, for such channels, after their abandonment by glacial waters, provided ready-made routes for later drainage. In many instances the present streams are very small in relation to the size of the valleys they occupy. The existence of misfit streams in many marginal channels has long been recognized and is to be expected in many subglacial ones also.

Some subglacial channels, however, are occupied by streams of considerable size as, for example, the valley of Limestone Creek between D11 and D26 (to be distinguished from the much larger Limestone Creek farther west). The creek rises at D6 and enters the channel at D11. A tributary rises at D8 and enters the channel at the head of the latter. The west-east part of the channel, running along the hillslope, was formed along the ice margin. Where the feature swings around to a northward direction, a steep bluff up to 20 feet high extends for a quarter of a mile along its eastern side and continues on the southern side of the channel. Presence of this bluff excludes the possibility that the eastward-flowing meltwaters that cut the marginal west-east part of the channel continued eastward beyond it and shows that the meltwaters turned to flow beneath the ice. A mile to the north two more marginal channels (D23) join the Limestone valley, indicating that their waters also followed the valley beneath the ice. A short distance farther north another marginal channel (D24) functioned in the same manner. Finally, the Limestone valley is joined by channel D25, which is an example of a marginal channel that turns abruptly through a right angle to become a subglacial chute. It thus appears that Limestone Creek between points D11 and D26 utilizes routes followed by glacial waters, and its course is therefore marked as a channel in Plate 1.

In view of the evidence of meltwater flow into the ice one must inquire at this point as to its subsequent history. In some instances meltwaters appear to have emerged from the ice at a point east of their point of entry. This is suggested by the fact that many marginal channels are of considerable size at their heads (*i.e.*, at their western ends). The ice-dammed lake hypothesis would explain this as a result of these channels having been lake overflows; this explanation is unsatisfactory since lacustrine deposits are generally absent. More probably these marginal channels have been formed by streams that flowed partly along the ice margin and partly within and beneath (and perhaps on) the ice.

In other instances subglacial chutes show that meltwaters plunged down the slopes of the land surface buried beneath the ice, in places descending as much as 200 feet. Deposits are absent at the lower ends of most of the chutes. This suggests that the streams that cut them continued to flow with undiminished velocity after becoming englacial. On the other hand, in a few areas, particularly along the middle part of spur D below the main limestone scarp, areas of sand and gravel, the latter predominating, appear to have been deposited by streams flowing beneath and into the ice. The ice-dammed lake hypothesis would explain these

deposits as having been laid down by streams flowing along the ice margin. The following evidence is against this view and favors the interpretation put forward here.

(1) The distribution of the deposits appears to be more closely related to the subglacial than to the marginal channels. For example, a small patch of sand and gravel occurs near the northern end of channel system D14, and a larger patch (D17) suggests deposition by the meltwaters that plunged down the limestone scarp beneath the ice margin in excavating channels D15 and D16.

(2) Where not subsequently dissected, the deposits commonly form low, irregular mounds that may be termed kames. These kames, however, do not have the steep ice-contact slopes and flattish surfaces that one might expect to find if they had been formed between the ice margin and the hillslopes.

(3) All the deposits are in places cut into by both marginal and subglacial channels. They must have been laid down therefore before either type of channel was cut at the points where dissection occurs. The deposits could not have been laid down between the ice edge and the hillslopes since subglacial dissection of them could not then have taken place. This suggests that deposition occurred, in part at least, at a time when the ice edge lay against or a little south of the steep, north-facing limestone scarp. Meltwaters flowed for short distances along the ice margin but then plunged down the scarp to flow beneath the ice, cutting some of the numerous gashes that now occur in the scarp. The materials produced by this erosion were partly redeposited farther down the slope beneath, and probably also partly within, the ice. As the ice continued to thin, and its edge continued to retreat down the slope, the waters flowing down the slope beneath the ice eroded to lower and lower levels, so that the deposits previously laid down by subglacial and englacial streams were partly dissected by the former. As the ice margin itself retreated across the deposits, further dissection occurred where waters flowed along it before themselves plunging beneath the ice.

Additional evidence of drainage beneath and through the ice is provided by the southern part of the valley of Cowaselon Creek. The valley is here about a mile wide and 400 feet deep. Its upper edge is generally a sharp convex break of slope, below which its sides descend steeply for some 200 to 250 feet. The central part of the valley, except for a strip along the course of the creek, is filled with deposits of sand and gravel, which in places extend up the steep valley sides. These deposits form numerous kames, ranging in height from a few feet to about 50 feet. The great majority of the kames are long and narrow and are aligned either straight up and down the valley sides or at a small angle with the line of steepest slope pointing slightly down-valley. Such kames extend to the head of the Cowaselon valley, about 2½ miles south of the southern limit of the area mapped in Plate 1.

Since the kames lie perpendicular to the direction of ice movement, they show that at the time they were formed a mass of dead ice occupied the Cowaselon valley. The alignment of the kames also indicates that they were deposited by supraglacial, englacial, and/or subglacial streams flowing toward the axis of the valley. The steep sides of many of the kames indicate that many of these streams were, in fact, subglacial. The abundance of kames on the western side of the valley north of D9 and their almost complete absence from the corresponding portion of the eastern side of the valley is largely the result of meltwaters from the west flowing into the valley.

At point D19 a deep subglacial chute, which continues a marginal channel of similar dimensions, runs straight down the valley side through a vertical distance of 300 feet. Along one side of the chute is a straight, steep-sided kame. This indicates that the waters flowing down the steep western side of the Cowaselon valley were at first engaged in deposition and built the kame in a steeply inclined subglacial tunnel. The meltwaters flowing through the tunnel later became erosive, probably owing to lowering of the englacial water table, and cut the subglacial chute, destroying part of the kame in the process.

A short distance to the north another subglacial chute (D20), which continues a short but well-defined marginal channel, descends part way down the valley side and is then continued by a kame. It appears that the subglacial waters changed from an agent of erosion to an agent of deposition at this point, probably because this was the level of the englacial water table at the time these features were formed.

Immediately to the north another mass of sand and gravel (D21) extends down the valley side, tapering from about 1000 feet wide at the top to about half this at the bottom. It was deposited by the waters that cut channel system D22. The main channel of this system runs straight down the side of the Cowaselon valley along the middle of the deposit, dividing it into

two steep-sided ridges. This indicates that, as in the case of the kame and channel at D19 described heretofore, deposition occurred in a steeply inclined subglacial tunnel, and this was followed by erosion by a subglacial stream flowing through the tunnel. The valley down the center of the depositional area (here interpreted as a subglacial chute) could hardly be of postglacial origin, since this would require a postglacial stream to have developed along the crest of a sand and gravel ridge running down the valley side.

Similar, but smaller, features to those at D21 occur at D18. Farther south, at D13, a kame runs straight down the valley side through a vertical distance of about 300 feet, and a subglacial chute follows one side of it. Still farther south, at D4, another kame extends through a similar vertical distance. This last kame does not appear to be related to a meltwater channel. Instead, it represents the deposits of a land stream that flowed into the ice occupying the Cowaselon valley, for it continues exactly the direction of the upper course of a small stream that enters the valley from the southwest, which stream changes direction at the beginning of the kame to wind its way down the valley side.

The kames of the Cowaselon valley are composed for the most part of coarse gravel with irregular thin beds of sand, except in the extreme southern part of the area studied, where the kames in the center of the valley are formed of silts and sands. The upper limit of deposition in this southern area is commonly very well defined at an altitude of approximately 900 feet. At this level most of the kames disappear, especially on the eastern side of the valley, where their crests show a marked flattening between 850 and 900 feet. Were these kames a little wider, their upper parts would form a kame terrace. It thus appears that during part of the time the dead ice in the Cowaselon valley was wasting away, a lake existed in the southern part of the valley at a level of about 900 feet. One might therefore anticipate the existence at the southern end of the valley of a channel at about 900 feet through which the lake waters escaped. A well-marked channel does exist there, but its floor level is slightly below 1100 feet. Since this channel floor is the lowest point in the drainage divide, the lake waters must have escaped northward, either along the ice margin or through the ice itself. A short narrow northward extension of the lake along the eastern margin of the ice occupying the valley is indicated by the flat-topped kames mentioned, but north of these kames for about 2 miles there is neither erosional nor depositional evidence at the appropriate altitude of marginally flowing meltwaters on the eastern side of the valley. It appears, therefore, that the meltwaters here flowed through the ice. Some of these waters then followed channels E1 and E2.

The phase of deposition up to the 900-foot level is the most clearly represented in the Cowaselon valley. Less well-defined phases occurred at slightly below 1100 feet, controlled by the channel at the southern end of the valley, and at about 750 feet, controlled by the channels at E4.

Thus, the Cowaselon valley evidence presents the following picture. As the ice margin retreated down the slopes of the spurs, a long tongue of ice remained in the valley. This tongue became thoroughly decayed, and water was able to flow freely through it. The level of the water in the ice was controlled by rock outlets, first at the southern end of the valley, and then on the eastern side of the valley toward its northern end. An englacial water table was thus developed. Where the surface of the ice lay below the level of this water table, a lake existed on top of the ice, as is demonstrated by the lacustrine deposits forming kames at the southern limit of the mapped area. The englacial water table also controlled the level of deposition within and beneath the ice: its level thus determined the upper limit of subglacial kame formation at any time and the lower limit of subglacial stream erosion. The lowering of the water table permitted the subglacial streams to dissect the long kames they had previously deposited on the western side of the valley.

It should perhaps be added that the retreat features of the Cowaselon valley are by no means uncommon; similar features have been described in Sweden (Mannerfelt, 1945) and Britain (Sissons, 1958). It may be noted also that Holmes (1952, p. 1008) concluded from studying the lithologic composition of a kame terrace in the Keeney valley, about 20 miles southwest of the Cowaselon valley, that upland drainage had carried material down against stagnant valley ice during deglaciation.

Although no definite stages in the retreat of the ice from the Syracuse-Oneida area before Great Lakes waters flowed through it can be recognized throughout the area, evidence indicates the general form of the ice margin. The evidence of the Cowaselon valley clearly shows that a narrow tongue of ice remained in that

valley after the higher ground on either side had become ice free. Kames and kame terraces in the Oneida valley show that a tongue of dead ice lay in that valley also. In the main valleys to the west such depositional evidence appears to be absent within the area studied but it is a reasonable presumption that they too were occupied by tongues of ice. This is implied by the marginal channels, which show that the ice margin closely followed the contours of the spur slopes and hence projected into the valleys. The same conclusion is indicated by the trend of the small end moraine on spur C.

This moraine appears to provide the only clear evidence that the ice was active at any time during its final retreat from the area. It consists of scattered mounds up to 30 feet high that appear to be composed mainly of till with beds of poorly stratified sand and gravel in places. In its eastern part (C2), angular limestone blocks up to 5 feet across lie on the morainic mounds or project from their grass-covered slopes. The apparent restriction of the moraine to spur C may be due to topographic factors, since it is possible that here an advance of the ice was retarded by the sharply rising ground immediately south of the moraine. Apart from this moraine the evidence suggests that the marginal zone of the ice was essentially stagnant during its final decay. Stagnation of the ice tongues in the deep valleys is expectable and in the Cowaselon and Oneida valleys is demonstrated by the ice-contact forms. Eskers near D5 and C1 indicate stagnation of the ice in the areas where they occur since they are not parallel with the direction of ice movement. In general, the preservation of the abundant evidence of subglacial drainage and the apparent absence of till moraines (apart from the exception noted) seem to suggest an essentially stagnant rather than an active ice-marginal zone.

Great Lakes Drainage

Most of the erosional and depositional features near Syracuse have been described by Fairchild, so only a brief description is necessary here. The principal channels, such as A3, A5, B6, C11, and C12, are 2–3 miles long, 100–150 feet deep, commonly have flattish floors 600–1200 feet across, and have steep sides which are commonly vertical where cut in limestone. They begin well above the levels of the valley bottoms to the west of them, and for the most part their floors slope down gently eastward. Toward the eastern end of channels A3, A5, and C12, abandoned waterfall sites occur, associated with plunge basins in channel C12. Numerous other waterfall sites, some of them associated with plunge basins, occur in a belt about 10 miles long and up to 1½ miles wide in the limestone region immediately south of channels A5, B6, and C11. Some of these are quite small, but waterfall sites at A4, B4, B5, and C10 are up to 150 feet high and occur at the heads of vertical-sided flat-bottomed gorges. Deposition by Great Lakes waters took place mainly in and near the south-north valleys (Onondaga, Butternut, and Limestone). The deposits, which are mainly coarse gravel (in some places composed largely of limestone boulders 1–2 feet across, with occasional larger blocks), have been subsequently dissected by slightly later meltwater erosion and by postglacial-stream erosion to form terraces bordering the south-north valleys and the channels.

According to Fairchild these features were formed by Great Lakes waters flowing along a retreating ice margin aligned southwest-northeast. This conclusion was unavoidable in view of his marginal-drainage interpretation. In his own words (1909, p. 36),

> "The production of cataracts on the west side of each valley . . .proves that the waters stood much lower in each valley, successively, than in the next valley on the west. Such discordance of the water-level could not have existed if the ice front had extended in an east and west direction across the ridges and valleys."

This view is difficult to reconcile, however, with the evidence already discussed indicating that prior to the establishment of Great Lakes drainage the ice margin was closely related to the contours of the hill slopes, tongues of ice extending up the major valleys. Fairchild overcame this difficulty in his final papers (1932a; 1932b) by suggesting that the features in the Syracuse area were formed by Great Lakes waters during a re-advance of the ice and the subsequent retreat. He thus regarded the features on the hillslopes south of channel C11 and east of the Limestone valley to the eastern limit of the area considered here as having been formed entirely by local waters during the penultimate retreat of the ice.

The problem of the orientation of the ice margin during the Great Lakes drainage through the Syracuse-Oneida area appears to be best approached by first ascertaining the upper limit of the region affected by this drainage. On spur A this limit is fixed by the edges of

channel A3 and by the presence of uneroded and partly eroded drumlins as lying between altitudes of 920 and 940 feet. On spur B channel B3 indicates that Great Lakes waters affected this area to just above 900 feet. On spur C the lip of the former waterfall at C9 is at 920 feet. Farther east on this spur large channels which are cut into the limestone scarp and the slope beneath it extend up to altitudes of between 900 and 950 feet. Outside the area considered here, on the spur to the west of the Onondaga valley, Fairchild (1909, p. 24) reported that the highest features cut by Great Lakes waters appeared to be between 920 and 940 feet. Thus over a west-east distance of about 18 miles there is evidence that the maximum altitude reached by Great Lakes waters when flowing through the area was between 900 and 950 feet. If, as seems probable, this evidence indicates the approximate position of the ice margin at the time Great Lakes waters last began to drain eastward, the eastward slope of the ice margin must have been very gentle (although not quite as gentle as these figures suggest, since allowance must be made for subsequent isostatic recovery). Such a gentle gradient seems more in accord with that to be expected along the margin of a large ice sheet than does the steep gradient required by Fairchild's hypothesis. It is also expectable because immediately before Great Lakes waters began to flow through the Syracuse-Oneida area the meltwater drainage was westward in the extreme west (A1 and A2), implying a westward slope of the ice margin there at that time. As the ice margin retreated prior to the Great Lakes drainage phase, the location of its highest point apparently shifted westward. When channels B1 and C3 were cut, the highest point of the ice margin lay between them, for these channels start at approximately the same altitude (1100 feet) but descend in opposite directions. By the time channels A2 and B2 were cut by waters flowing in opposite directions, the highest point of the ice margin lay farther west (between spurs A and B) and was about 100 feet lower, for these channels start at around 1000 feet. Further retreat, possibly accompanied by a further westward shifting of the highest point of the ice margin, allowed eastward drainage of Great Lakes waters to be established.

The highest feature produced by Great Lakes drainage on spur D is channel D27. The floor of this channel at its head is at an altitude of about 810 feet, and its downcutting probably commenced between 880 and 900 feet. This channel is far larger than any of the channels on the slopes above it and cannot be ascribed to local meltwater drainage. It also differs from these smaller channels in its length, for it extends for 6 miles to the Cowaselon valley. For the last mile and a half of its course it is a steep-sided trench about 100 feet deep, with a flat floor 300 to 500 feet across cut in shale.

On spur E the highest feature related to Great Lakes waters is channel E5. This channel is much larger than the channels at higher levels cut by local waters, being 100 or more feet deep for most of its course, and it also differs from them in that it forms a continuous feature from one side of the spur to the other. It is continued across the Oneida valley by an almost flat-topped mass of gravel (E6) whose surface lies between 630 and 650 feet. Thus it appears that at the time Great Lakes waters began to escape through the Syracuse-Oneida area the altitude of the ice margin declined gently eastward in the western half of the area, lying, in terms of present altitudes, between 950 and 900 feet over a distance of 18 miles, and declined somewhat more steeply in the eastern half of the area from around 950–900 feet to about 650 feet in a distance of 12 miles.

The failure of Fairchild and others to recognize that Great Lakes waters flowed through channels D27 and E5 appears to be due to two related groups of facts. First, these two channels, although far larger than the channels at higher levels on the same spurs, are, nevertheless, smaller than the vast channels cut by Great Lakes waters near Syracuse. This is obviously difficult to explain by the marginal-drainage hypothesis. Secondly, successive sequences of large channels and related features do not occur on spurs D and E below the level of channels D27 and E5. Both these groups of facts are readily explained on the interpretation put forward here, for they resulted from increasing quantities of Great Lakes waters having found their way through and beneath the ice.

Clear evidence of this is provided by a group of channels on spur C. Channel C7 is a great trench up to 150 feet deep and 700 feet across running straight down the steep western side of the Chittenango valley through a vertical distance of about 400 feet. The present tiny stream could not have formed this feature. It is a great subglacial chute cut by some of the Great Lakes waters plunging beneath the ice. The various routes followed by these waters along the ice margin before they turned to flow

beneath the ice are indicated by the channels that join channel C7 from the northwest. A short distance to the southeast are two smaller subglacial chutes (C5 and C6) whose origin is shown by their relationship to marginal channels. A short distance northwest of channel C7 a large marginal channel (C8) 600 feet broad and 50 to 100 feet deep follows the scarp face for a mile. The channel then suddenly turns through a right angle and runs straight down the hillside as a subglacial chute, descending 200 feet in three-quarters of a mile. Other large channels that carried Great Lakes waters first along the ice margin and then beneath the ice occur to the west-northwest. The westernmost feature of this group is channel C10, which, for the first half mile of its course, is 100 feet deep and 600 feet broad with vertical sides. The orientation of this part of the channel suggests it was cut along the ice margin, but it appears that the waters then flowed beneath the ice, for the channel changes direction and descends abruptly toward the floor of channel C11. Evidence of flow beneath the ice is also provided by waterfall site C9. This waterfall site is a vertical-walled semicircular recess about 600 feet across and 130 feet high cut into the limestone scarp, at the top of which is a water-worn limestone platform. For these features to have been formed, the ice must have lain against the limestone scarp, and its edge must have been at least level with the top of the scarp. Since marginal channels on spur C consistently indicate that the ice margin followed the scarp edge, the waters that cut the recess must have done so as they plunged down between the scarp face and a wall of ice, subsequently flowing beneath the ice.

The features on spur C described in the preceding paragraph resemble those produced by local waters on spur D except that they are on a much larger scale. They are repeated on an even larger scale on spurs A and B, south and southeast of Syracuse. The channels here fall into two distinct groups. One group, represented by channels A3, A4, and A5, B4, and B6, as well as by a large number of smaller channels, trends approximately west-east. The other group, represented by channel B5, the Butternut valley north of its junction with channel A3, the Onondaga valley, part of the Limestone valley, and a number of smaller features, trends more nearly south-north. Fairchild's southwest-northeast ice margins appear to represent a compromise between the directions of these two groups of features and do not permit either group to be explained satisfactorily.

Among the difficulties encountered by Fairchild's interpretation are the following.

(1) He concluded that channels A5, B6, and C12 formed a continuous route for the glacial waters at one stage in the retreat, and that at a later stage these waters followed channel A5 and the Butternut valley. That the glacial waters followed these routes is demonstrated by both channels and terraces. But that they did so along the ice margin in both cases is difficult to believe. According to the marginal-drainage view, the edge of the ice could not have been more than a mile north of channels A5, B6, and C12 at the time they functioned together, since otherwise the waters would have followed a route farther north. Therefore the ice edge must have trended west-east across spurs A and B. But by the time channel A5 and the Butternut valley carried the glacial waters, the ice margin must have changed direction to trend southwest-northeast. The ice margin must have therefore remained more or less constant in position at the head of channel A5 to allow that channel to continue functioning while only a short distance away it retreated 4 or 5 miles from near the lower end of channel C12 to some point west of the Butternut valley. Such an occurrence seems highly improbable.

(2) As already noted, the highest altitude affected by Great Lakes waters on spur A is between 920 and 940 feet, this being the maximum altitude of the edges of channel A3. The floor of this channel at its head is just below 800 feet. The maximum altitude of the water-swept limestone region around channel A4 is also just below 800 feet. Abundant evidence of meltwater erosion occurs below this altitude down to 550 feet, which is the altitude of the floors of channels A5 and A10 at their heads. This evidence is consistent with these features having been formed along an ice margin as Fairchild maintained. But it indicates that the ice margin trended, not southwest-northeast, as he concluded, but approximately west-east.

(3) On spur B approximately west–east-trending ice-margin positions are demonstrated by channels B4 and B6 and by the numerous smaller features between them. But if the ice margin trended west and east across spurs A and B, channel B5 and the part of the Butternut valley between the western ends of channels A3 and A5 could not also have carried waters along the ice margin, for they lie almost per-

pendicular to and cross the marginal features on spurs A and B.

(4) If, as Fairchild maintained, the glacial waters flowed only along the ice margin there should be continuous sequences of erosional and/or depositional features down the slopes of the spurs crossed by Great Lakes waters. On spurs B and C, however, gaps occur in the sequences. The floor of channel B6 at its head is at 530 feet. A mile to the north is a col (B7) at 510 feet. According to Fairchild's hypothesis, this col should have carried Great Lakes waters after they ceased to flow through channel B6. That it did not do so, however, is indicated by the uneroded drumlins on its southern slope, by the absence of sand and gravel deposits from its immediate vicinity, and by its somewhat irregular floor and rather subdued character compared with other channels on the same spur. It seems probable that the col is a meltwater channel, but that it was formed prior to the last retreat of the ice from the area. On spur C there is a gap in the sequence between channel C10, commencing at 840 feet, and channels C11 and C12, the upper limits of whose sides are below 700 feet.

There is thus abundant evidence of marginal drainage on parts of spurs A and B, as recognized by Fairchild, although the ice margin trended approximately west-east, not southwest-northeast as he concluded. On other parts of these spurs the marginal hypothesis is inadequate, however. Here the approximately south-north drainage evidenced by certain channels and deposits points to drainage beneath and through the ice. It is only by postulating such drainage that the apparently contradictory evidence of gentle ice-margin gradients and great waterfall sites in the same area can be reconciled. The great embayment about half way along the south side of channel A5 is walled by a cliff 150 feet high at the top of which is an area of water-worn limestone, with deep gullies and underground passages along joint and bedding planes that carried waters to the cliff face where they plunged down in great falls. These falls could have existed only if the ice margin was situated immediately to the north, and they must have developed between the limestone cliff and a wall of ice. Similar evidence points to a similar origin for channel B5, and smaller waterfall sites, some of them with plunge basins at their bases, occur along the scarp face forming the southern wall of channel B6 between channel B5 and the Butternut valley.

The waters that followed these and other routes probably flowed beneath the ice initially. In view of the great volumes of water involved and the probably limited thickness of the ice, however, the subglacial interpretation appears unsatisfactory as a complete explanation. Furthermore, it is clearly invalid in the case of the Butternut valley north of its junction with channel A5, for here extensive outwash terraces occur. Since, as pointed out, it is highly improbable that this valley carried meltwaters along the ice margin, it appears that the only satisfactory alternative is that they followed this valley in a channel walled by ice. It seems likely that flow along the valley was initiated subglacially, the roof of the tunnel subsequently collapsing to produce an open trench through the ice. The abandonment of channel A5 for the route northward down the Onondaga valley to Syracuse probably took place in a similar manner. On a smaller scale, the terraces in the Butternut valley near the eastern end of channel A3 and in the Cowaselon valley near the eastern end of the long channel D27 indicate that the flow of Great Lakes waters into these valleys caused the tongues of dead ice extending up them to be severed. The best example occurs in the extreme eastern part of the area, in the Oneida valley. Here a large mass of sand and gravel (E6) forms a feature that resembles a great dam built part way across the valley. The mass extends west-east for a mile, has an almost flat top a quarter of a mile broad that inclines very gently eastward, and is bounded by steep slopes up to 100 feet high. The feature continues exactly along the direction of channel E5, and it clearly represents the infilling of a deep trench through the ice, its steep slopes to north and south being ice-contact slopes: thus Great Lakes waters cut straight across the dead ice tongue occupying the Oneida valley.

In summary, it appears that during the final retreat of the ice from the Syracuse-Oneida area, eastward drainage from the Great Lakes was initiated at a time when the ice edge declined in altitude from between 950 and 900 feet in the western half of the area to about 650 feet in the extreme east (in terms of present altitudes). This ice mass was already much decayed in its marginal zone, as indicated by the abundant evidence of subglacial and englacial drainage developed before the establishment of Great Lakes drainage. The earliest Great Lakes flow, however, is represented by long marginal channels, such as D27 and E5, the lengths of which contrast with the shortness of adjacent,

slightly higher channels, produced by locally derived waters. The probable explanation of this is that the openings in the ice developed by local waters prior to the Great Lakes flow were incapable of absorbing this flow immediately, so that an important marginal drainage was established. The subsequent history of the Great Lakes drainage is largely one of the successive abandonment of marginal routes for others leading more directly to the low ground in the north. A short cut across the ice tongue in the Oneida valley was rapidly established by the waters that had passed through channel E5. Subsequently the waters followed a route slightly farther north, laying down the kame terraces and associated deposits around E8. The waters then abandoned spur E and flowed from channel D27 northward along the northern part of the Cowaselon valley, cutting off the remnant of the dead ice tongue in that valley (as indicated by flat-topped kames according in level with the terrace around the valley mouth) and also separating from the main ice mass an area of dead ice along the northern foot of spur E (part of which on its subsequent decay produced the kame and kettle topography south of Oneida). The volume of water flowing through the long channel D27 decreased as the series of large subglacial channels on spur C developed until, eventually, flow across spur D ceased, the last phase of flow across this spur being recorded by the kames with steep, north-facing ice-contact slopes extending from D30 to D28 and by channel D29, whose floor is at approximately the same level as the nearly accordant summits of the kames. Similar developments took place farther west, the major routes by which Great Lakes waters took short cuts to the lower ground, detaching portions of the marginal zone from the main ice mass as they did so, being represented by channel C12 and the northern sections of the Butternut and Onondaga valleys.

Former Lakes

Only the highest shore features of Lake Iroquois are shown in Plate 1. They increase in altitude eastward, as described by Fairchild (1909; 1916), from about 430–440 feet at the northern end of the Butternut valley to about 460 feet around Oneida. Above the Iroquois shore features, numerous very small, but nevertheless distinct, shore features occur east of the Chittenango valley, extending up to a maximum altitude of about 515 feet. As many as nine faint steps can be distinguished in places in a vertical interval of 40 feet. Bars are small and uncommon and occur only where abundant unconsolidated material was available from kame areas as at D31 and 32. These kame areas are cliffed in places, the cliff base being at about 500 feet. Subdued mounds of sand and gravel that occur below this altitude near the kame areas (above the level affected by Lake Iroquois) appear to be wave-modified kames.

The outlet (or outlets) of the lake in which these features formed was presumably east of Oneida. The absence of evidence of meltwater erosion related to the last retreat of the ice below an altitude of about 500 feet east of the Chittenango valley indicates that the lake existed there while the ice withdrew. To the south and east of C13 a large area of sand and gravel, delta-bedded in places, has a flat top at about 500 feet. The position and bedding of this deposit show that it was laid down by meltwaters that had come through channel C12. Therefore this channel originally discharged into the lake. These sand and gravel deposits are, however, dissected by large channels. Furthermore, the floor of channel C12 northeast of the abandoned waterfall site is below 420 feet in altitude. The lake must have been drained, therefore, while channel C12 was being used by Great Lakes waters. Considerably later, submergence to altitudes up to 460 feet occurred owing to encroachment of Lake Iroquois as a result of isostatic tilting.

Fairchild's Syracuse-Oneida Moraine

Fairchild (1932b) described an end moraine extending from Syracuse to Oneida along the hill-foot zone and continuing both west and east beyond the limits of the area considered here. The moraine as marked on Fairchild's map (1932b, opp. p. 627) includes the kames and kame terraces south of Oneida, the kames along the northern end of spur D, the outwash terraces at the northern ends of the Cowaselon, Limestone and Butternut valleys, and the dissected outwash south and east of C13. Most of these deposits cannot legitimately be described as "moraine", and their relationships to each other and to the meltwater channels show that they were not formed contemporaneously. Since there appears to be no other evidence of end moraines along this belt it must be concluded that the Syracuse-Oneida moraine does not exist.

Evidence of More Than One Period of Great Lakes Drainage

Evidence indicates that Great Lakes waters flowed through the Syracuse-Oneida area on at least two occasions. Channel E7 appears to have been cut by these waters, for it is a quarter of a mile broad, its northern side is up to 70 feet high, and its southern side is up to 160 feet high. Its eastern end, however, is almost blocked by a mass of sand and gravel forming kames and kame terraces that extend up to 60 feet above the channel floor. These deposits show that channel E7 was not formed during the last retreat of the ice from the area. The floor altitude of channel D34 suggests that it was formed contemporaneously with channel E7. That channel D34 was not formed during the last retreat of the ice is also suggested by the absence of related outwash deposits from its eastern end.

Channel D33 is 4½ miles long and ranges in width between a quarter of a mile and half a mile. The eastern end of this channel is blocked by a mass of sand whose nearly flat top, about 50 feet above the channel floor, is bounded by well-defined, sinuous ice-contact slopes, indicating that dead ice occupied the channel at some time after its formation by Great Lakes waters. The same conclusion is indicated for channel C11 by the low kames and the sand and gravel terrace that obscure the first half mile of its course.

Reasons have been given for concluding that channel B7 was not formed during the last retreat of the ice. Its position and altitude suggest that it carried Great Lakes waters that had passed through channel A10, which, along with channel A7, also appears to have been formed prior to the last retreat. Farther south on spur A the great loop channel on the south side of channel A3 was apparently partly occupied by ice when the west-east part of channel A3 was carrying Great Lakes waters during the last retreat. The entrance to the loop is occupied by a mass of sand and gravel up to 50 feet thick, which has an irregular boundary on its southern side suggesting an ice-contact slope. A detached mass of dead ice in the loop would explain the sudden termination of the sand and gravel deposit on its southern side and would also account for the marshy area, suggestive of a kettle, immediately south of the deposit. In the eastern part of the loop other sand and gravel deposits form mounds and a small terrace. These mounds appear to be kames formed in association with dead ice, and not the result of dissection of an outwash mass by the present stream, for a section in one of them shows that the dip of the beds accords approximately with the slopes of the feature.

There is thus adequate evidence that some of the great channels in the Syracuse-Oneida area were occupied by dead ice during the final deglaciation. It is possible to argue that this resulted from frequent minor oscillations of the ice margin during the final retreat, so that after a certain channel was formed the ice advanced again and occupied it, subsequently becoming stagnant therein. This explanation seems inadequate, however, since, apart from the complex ice-margin oscillations it would necessitate, it does not accord with the pattern of retreat established from the marginal, subglacial, and englacial drainage evidence discussed heretofore, and the numerous moraines that one might expect to have been formed at the limits of at least some of the readvances do not exist. It appears, then, that the evidence indicates at least two periods of Great Lakes drainage through the area.

In addition to the channels just discussed, certain other major channels, which did carry Great Lakes waters during the last deglaciation of the area, were formed for the most part at an earlier date. The waterfall sites along the southern walls of channels B6 and A5 indicate that these channels were already in existence before the last retreat and were occupied by ice. At the bottom of the waterfall site on the southern side of channel A5 is a deposit of sand and gravel containing limestone blocks up to 10 feet across. It was laid down at the time the waterfalls functioned and was not swept away subsequently by the waters flowing through A5 since it lies in the embayment on the southern side of the channel. The level of the bedrock surface beneath the deposit is approximately the same as the level of the bedrock floor of channel A5. This implies that this channel had been excavated to its present depth before the last retreat of the ice. That channel B6 was excavated to its present depth prior to the last retreat is suggested by the accordant junction with it of channel B5. Since the relative altitudes of channels A5, B6, and C12 show that they functioned at the same time, it follows that C12 (at least above the waterfall site) was also excavated to its present depth prior to the last retreat. The same conclusion applies to channel A3, since the level of its bedrock floor is approximately the same as that of the bed-

rock surface beneath the sands and gravels in the big loop channel on its southern side.

Thus, four great channels in the Syracuse area, which were utilized by Great Lakes waters during the final deglaciation, as terraced outwash masses prove, appear to have been excavated to their present depths at some time prior to the last retreat. The main erosional effects of the Great Lakes waters in these channels during the last retreat were the addition or modification of the embayments due to waterfalls on the southern sides of A5 and B6, and the modification of the waterfall sites at the eastern ends of A3, A5, and C12.

The drumlins near Syracuse suggest that two periods of Great Lakes drainage occurred during the oscillatory retreat from the Wisconsin maximum. In the small area north of channel A5 and west of the Butternut valley are about 50 drumlins. Many of these drumlins are 100–150 feet high, 2000–3000 feet long, and 700–1000 feet broad. Within the same area, and commonly lying between the big drumlins, are smaller drumlins, which are generally between 10 and 40 feet high, 800 to 2500 feet long, and 200 to 300 feet broad. Thus the smaller drumlins are long, low, and narrow compared with the others. If, as appears to be generally accepted, the form and size of drumlins is related, in part, to the intensity of former ice movement, it would seem that two distinct periods of drumlin formation occurred in this area, the larger drumlins having been formed during the earlier period. This is also suggested by the existence on several of the bigger drumlins of long, low, narrow tails of dimensions similar to those of the smaller drumlins.

The final utilization of the large channel A7 by Great Lakes waters postdates the period of formation of the large drumlins, since the channel cuts cleanly across two of them. It predates the formation of the small drumlins since two of them (A8) extend a short distance onto the channel floor, and their rounded forms are unmodified, whereas a third unmodified small drumlin (A6) extends into the area that Great Lakes waters probably flowed through before they plunged over the limestone cliff at the head of channel A7. That the channel was not utilized by glacial waters during the final deglaciation is also indicated by its slightly irregular, partly till-covered floor, the small, partly obscured, plunge basins at its head, and the reverse slope at its eastern end due to outwash deposited by waters flowing through channel A5.

Since it seems highly improbable that the large drumlins have survived from a pre-Wisconsin glaciation, it appears that the evidence associated with channel A7 indicates two phases of Great Lakes drainage through the Syracuse area during Wisconsin times. This accords with Fairchild's conclusion (1932a; 1932b) which is accepted by Flint (1953; 1956; 1957). In terms of Flint's suggested correlations the small drumlins may have been produced by ice advancing to the Valders maximum, one period of Great Lakes drainage through the area having taken place during the advance to that maximum, the other during the retreat from it. If only these two periods of Great Lakes drainage through the Syracuse-Oneida area occurred, then the preceding retreat (culminating in the Ontario basin equivalent of the Bowmanville low-water phase) did not result in Great Lakes drainage through the area, being associated only with local drainage as Fairchild concluded (1932a; 1932b). The evidence for this conclusion of Fairchild lies outside the area considered here and cannot be discussed therefore. It may be remarked, however, that his conclusion may require modification in the light of modern ideas on glacial drainage. For example, his interpretation of the low-level channels (*e.g.*, Fairport-Lyons and Victor-Phelps) as marginal features encounters many difficulties, and there are good reasons for believing that these channels were formed by Great Lakes waters flowing across the dead marginal zone of the ice itself between walls of ice.

Whether or not only two periods of Great Lakes drainage through the Syracuse-Oneida area occurred in Wisconsin times, this writer suggests that the vast channel system has a much longer and more complex history than the preceding discussion might appear to indicate. Firstly, it seems reasonable to presume that the Great Lakes drained through the area in pre-Wisconsin times, for the ice borders appear to have been broadly similar during the different glaciations. Secondly, there is evidence of Great Lakes drainage before the formation of the larger drumlins. Thus the drumlin that forms part of the southern side of channel B7 belongs to this group, indicating that this channel was formed before these drumlins. Channel A10 was formed before the smaller group of drumlins, since two of these ascend its southern slope from the level of its floor (A9). This channel also appears to have been formed before the larger drumlins were produced (although it may have been utilized later), for the distribu-

tion of these drumlins appears to be partly controlled by it: to the south of the channel these drumlins commence exactly at its lip, although none of them have been eroded by the waters that cut the channel. Thus channels A10 and B7 both suggest Great Lakes drainage before the larger drumlins were formed, and these indications are mutually consistent, for the relative position and altitude of the two channels strongly suggest that they functioned together. Finally, the author finds it difficult to believe that the tens of miles of large channels could have been produced by only two periods of Great Lakes flow through the area, bearing in mind the rate of retreat of Niagara Falls, which is of dimensions comparable with many of the falls that formerly existed in the Syracuse area, and the limited time available, as indicated by carbon-14 dates. In the Syracuse-Oneida area alone there are nearly 40 miles of channels about 100 or more feet deep and averaging about a quarter of a mile wide, produced by Great Lakes waters, besides many smaller channels cut by these waters. The total length of channels of these dimensions cut by eastward-draining Great Lakes waters in the whole of central New York is greatly in excess of this figure. It has been shown heretofore that the last period of Great Lakes drainage did not utilize many of the large channels, and it produced only limited modifications of several other great channels. This means that, if only two periods of Great Lakes drainage through the area are postulated, almost all the vast channel system was produced during the earlier period. It seems much more likely that Great Lakes waters have flowed through this area on a number of occasions, each occasion after the first resulting in limited additions and modifications, until eventually the present vast channel system was produced.

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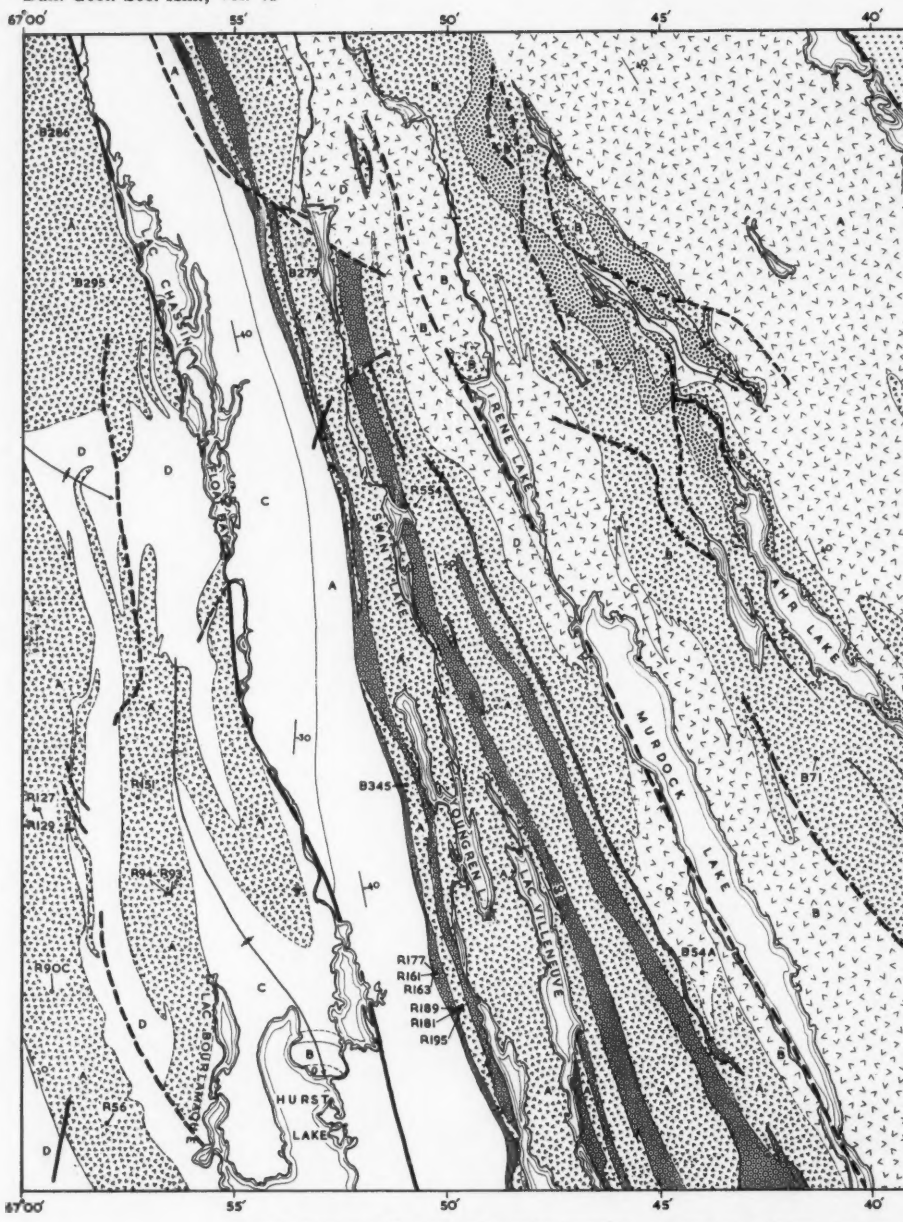
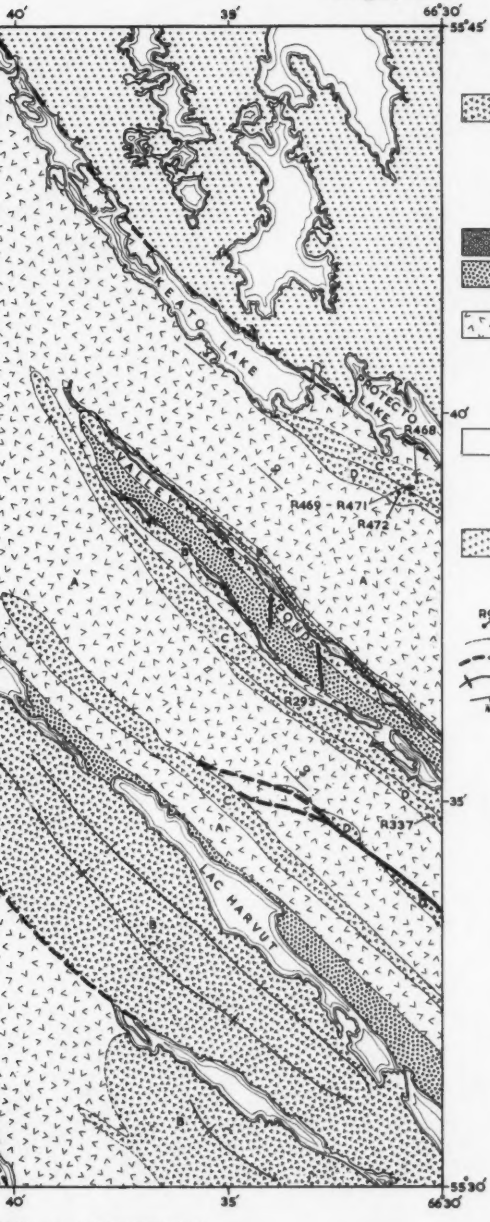


GEOLOGIC MAP OF THE AHR LAKE AREA

Baragar, Pl. 1



AHR LAKE AREA

LEGEND

GABBRO AND METAGABBROS
A. GABBRO, MINOR SEDIMENTARY ROCKS BETWEEN SILLS.
B. METAGABBRO, MINOR SEDIMENTARY ROCKS BETWEEN SILLS.
C. LEUCO-METAGABBRO
D. MELANO-METAGABBRO

LEOPARD ROCK

ULTRAMAFIC ROCKS. AGE RELATIVE TO UNITS ABOVE UNCERTAIN.

VOLCANIC ROCKS LARGELY
A. DOUBLET METABASALTS MINOR INTERBEDDED SEDIMENTARY ROCKS
B. MURDOCK PYROCLASTIC ROCKS
C. IRON FORMATION
D. HOWSE GROUP BASALTS

SEDIMENTARY ROCKS LARGELY
A. HOWSE GROUP SHALES AND WACKES
B. IRON FORMATION
C. DOLOMITE
D. QUARTZ WACKE, SHALE, AND BASALT

BIOTITE SCHIST. AGE RELATIVE TO UNITS ABOVE UNCERTAIN

R94 SPECIMEN LOCALITY
CONTACT (ASSUMED, ESTABLISHED)
FAULT (ASSUMED, ESTABLISHED)
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REA, QUEBEC

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PETROLOGY OF BASALTIC ROCKS IN PART OF THE LABRADOR TROUGH

By W. Robert A. Baragar

Abstract

Basalts and sedimentary rocks in the eastern half of the Labrador trough are profusely intruded by gabbro sills. In the Ahr Lake area, Quebec, where the igneous rocks were studied in detail, the gabbros are divided into normal gabbros, leopard rocks, and metagabbros. Metagabbros, which are confined to the eastern part of the area, are equivalent to normal gabbros except for two unusual varieties, leuco- and melanometagabbros.

Normal gabbro sills have differentiated into olivine-bearing lower, and pegmatitic upper, portions. The pegmatite of one sill (albite gabbro pegmatite) has sodic plagioclase in contrast with andesine in other pegmatites and a mesostasis interpreted as devitrified glass. Glass formation is attributed to volatile deficiency in the magma, but development of sodic plagioclase may have been due to later metamorphism facilitated by the presence of glass.

Leopard rock is a coarse-grained feldspathic gabbro spotted with 6- to 15-cm aggregates of altered plagioclase. Sills of leopard rock have narrow zones of medium-grained, sparsely prophyritic gabbro along their margins, and some are composite with intrusions of normal gabbro along their centers. Leopard rock is probably equivalent to normal gabbro with a high concentration of plagioclase, and it may have formed by partial melting of the mantle at high pressures or under conditions of moderately high water pressure. Magma with a load of suspended plagioclase may assume a flow structure upon intrusion such that plagioclase is concentrated in a central sheet and the margins are free of plagioclase clusters. Thus the marginal zones with few feldspathic clots are explained.

Melano- and leuco metagabbros occupy the lower and upper parts respectively of two sills and are separated by a narrow gradational zone. Each also occurs separately. Melanometagabbro is predominantly amphibole but shows relic olivine structures; leucometagabbro is predominantly altered plagioclase. Sills of melano- and leucometagabbros cannot be due to normal differentiation, and their origin is uncertain.

Petrographic, mineralogical, spectrographic, and chemical data for gabbros and basalts are given. Normal gabbros are tholeiites which show an iron-rich trend similar to the Skaergaard trend but with negligible alkali enrichment. Basalts of the area are equivalent to the parent magma of the normal gabbros, and differentiation was *in situ*. Low potassium and strontium are characteristic of the province. Trace-element behavior is similar to that in other tholeiitic provinces.

Metamorphism decreases westerly across the Ahr Lake area from quartz-albite-epidote-biotite subfacies of the greenschist facies through quartz-albite-muscovite-chlorite subfacies to subgreenschist facies rocks. Local development of pumpellyite and prehnite in the latter may indicate the zeolite facies.

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INTRODUCTION

General Statement

This paper is a study of a major basaltic province in northwest Labrador and northern Quebec. The province includes both sills and lavas and constitutes part of the Labrador trough (Harrison, 1952). Detailed study is confined to the Ahr Lake map area of northern Quebec (Baragar, 1958), but since this area lies across the igneous section of the trough the results may apply to the province as a whole. These basalts are associated with the geosynclinal phase of mountain building, but unlike the mafic rocks of many orogenic belts part of them has survived with only minor alteration. Thus the area affords the unusual opportunity of studying the original basalts of a geosyncline and comparing them with nongeosynclinal basaltic provinces elsewhere.

Historical Outline

Until the recent development of the New Quebec-Labrador iron-ore bodies, what little was known of the geology of northern Quebec was provided almost entirely by the remarkable explorations of Low from 1892 to 1900 (1896; 1898; 1901; 1902). Low's work included reconnaissance surveys of the Quebec coast from Ungava Bay to James Bay and several

overland treks into the Ungava interior. On one field excursion he traveled through Quebec from the St. Lawrence river to Ungava Bay and from Lake Melville on the Labrador coast into the Ungava interior. On this expedition he discovered the slightly folded rocks of the Labrador trough and its iron formation.

In 1929 the first ore body was discovered in the iron formation near the present town of Schefferville (Knob Lake). By the outbreak of war the discovery of several more ore bodies had awaken interest in the possibilities of eventual exploitation of the iron and stimulated a more intensive search. With the exception of the early war years, exploration lasted throughout the decade, and by the fall of 1950 enough ore had been proved to make a railway feasible connecting the iron fields with a port at Seven Islands. The railway was completed and the first ore shipped in 1954.

Meanwhile systematic geological mapping started by exploration companies was aided and extended by regional mapping programs of the Geological Survey of Canada and the Quebec Department of Mines. At present mapping of the trough is well advanced, but beyond its boundaries knowledge of the geology remains sketchy.

Methods

All mineral determinations were by optical means. Refractive indices were measured in oils using sodium light, and after each measurement the portion of the oil used was checked on a Leitz-Jelley refractometer. Precision in most cases is 0.002. Optic angles in augite were measured on a Leitz 4-axis stage from one optic axis to the other. Repeated measurement on the same grain indicate an accuracy of $\pm 1°$, but measurements on several grains in the same section commonly span a range of 1°–4°. The average is taken in each case. In olivine and orthopyroxene optic angles were generally determined indirectly from stereonet plots of one optic axis and the acute bisectrix.

Hess's (1949, p. 634) correlation curves based on β index and 2V were used for augite. Modifications proposed by Muir (1951, p. 713) are mainly in the ferroaugite field and have little effect on augites of the present study. Brown (1957, p. 537) has offered a new set of curves for the magnesian augites based on a restudy of Skaergaard augites. For the same point on the pyroxene diagram, Brown's curves give a 2V 7° lower and β index 0.005 higher than Hess's curves. The differences are attributed to: (1) failure to correct for exsolved lime-poor pyroxene in Hess's diagram, and (2) high titanium content in octahedral positions of Skaergaard augites (Brown, 1957, p. 536–537). Brown (1957, p. 536) suggests adding 3° to 2V before applying to Hess's curves, to offset the effect of lime-poor pyroxene on the curves. In view of the high titanium content of Skaergaard augites, Hess's curves were thought better suited for the present study. The 3-degree correction was not used, since the Ahr Lake course of crystallization is compared with the normal trend shown by Hess (1949, p. 634). A better estimate of composition may be obtained by applying the correction.

Correlation curves used for orthopyroxene are those of Hess (1952, p. 180), for pigeonite Hess (1949, p. 643), and for olivine Poldervaart (1950, p. 1073).

Plagioclase was determined mainly by measurement of α' on cleavage fragments, and Tsuboi's (1923, Pl. 1) correlation curves were used. More recent data are available (Smith, 1957, p. 217), but when α indices from which Tsuboi's curves were calculated are compared with Smith's α indices at corresponding plagioclase compositions they agree within 0.002, the error of measurement. Recent work (Smith, 1956, p. 189–190; 1957, p. 216–217) shows that α varies little from low- to high-temperature plagioclase; hence α' is probably a fairly reliable measure of composition. Where index determinations were unreliable because of alteration, plagioclase was determined on the universal stage using Federov and Nikitin methods and correlation curves of van der Kaaden (1951, Utrecht Univ. Thesis 17) and Turner (1947), respectively.

The β refractive index of actinolite was measured on suitably oriented crystals cut from thin sections or found in powders, and its specific gravity was determined by pycnometer. Winchell and Winchell's (1951, p. 433) correlation curves for the tremolite-ferrotremolite series were used.

Volumetric analyses are by the Chayes point-counter method and are based on 1000–2500 points per analysis, depending on the number and size of thin sections used. In most cases they are estimates only as alteration commonly obscures the boundaries between crystals. The alteration product was usually counted as the fresh mineral; for example, saussurite was counted as plagioclase, but this depends to some extent on judgment which in turn affects the precision of the analyses.

Acknowledgments

This study is based on field mapping done for the Geological Survey of Canada during the summers of 1954–1955. The laboratory investigation was carried out under the direction of Dr. Arie Poldervaart who was a constant source of inspiration, encouragement, and guidance. The writer is grateful for the opportunity to conduct the study at Columbia and Cambridge universities and wishes to thank staff members of both institutions for the considerable help he has received. Grateful acknowledgment is made particularly to the following: Prof. C. E. Tilley, Dr. S. R. Nockolds, and Dr. S. O. Argell for giving freely of their time and knowledge and contributing many valuable suggestions; J. H. Scoon for helping the writer with chemical analyses; and R. Allen who did the spectrographic analyses at Cambridge. Dr. K. K. Turekian guided the writer in spectrographic work done at Lamont Geological Observatory.

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Regional Geology

The Labrador trough is composed of sedimentary, volcanic, and intrusive rocks and extends from northwestern Labrador to Ungava Bay, a distance of about 550 miles (Fig. 1). It is about 60 miles wide in a section through Knob Lake but diminishes in width toward its northern and southern extremities.

The principal features of the regional geology are shown in Figure 2. The slightly to moderately altered rocks of the Labrador trough are bounded on each side by gneisses or schists. On the western side, rocks of the trough lie uncomformably on gneisses and dip gently eastward (Harrison, 1952; p. 7; Fahrig, 1956a; Bérard, 1957, p. 2); on the eastern side they dip steeply eastward and are in fault contact with schists and gneisses (Fahrig, 1951; 1956b). Within the trough the lithology changes from predominantly sedimentary on the western side to predominantly igneous on the eastern side; this is reflected topographically in a change from rounded hills and broad valleys in the western portion to a distinctly corrugated landscape in the eastern portion. Shale, quartzite, iron formation, arkose, and dolomite are the major components of the western assemblage, whereas shales, wackes, pillow lavas, pyroclastic rocks, gabbros, and ultramafic rocks are the main constituents of the eastern group. Typical assemblages from each portion are given in Table 1 for comparison. (*Cf.*, Roscoe, 1957, p. 4–5.) The Knob Lake section is taken from Harrison (1952) and represents the stratigraphic succession in the vicinity of Knob Lake. The Ahr Lake section is from the Ahr Lake area mapped by the writer (1958). It does not include the considerable thickness of gabbro and ultramafic sills which intrude rocks of the area.

Northeast-dipping thrusts are numerous within the trough and cause much repetition of stratigraphic sequences. The major thrust faults tend to separate rocks of the trough into discrete slices with different lithology, between which correlation is difficult. In a section across the trough near Knob Lake a western slice, 25 miles wide, contains most of the iron formation, dolomite, and quartzite; a central slice, 15 miles wide, has most of the gabbro and minor iron formation; and an eastern slice, 20 miles wide, contains most of the volcanic rocks and all the ultramafic sills. Within each slice are smaller faults of less significant displacement.[1]

Folds ranging from isoclinal to broad open structures are generally overturned to the southwest. Cross sections accompanying Roscoe's map (1957) of the Cambrian Lake area illustrate folding across the trough. Fold axes are horizontal or plunge gently northwest or southeast.

The intensity of metamorphism increases from west to east across the trough (Fahrig, 1957, p. 115; Bergeron, 1957b, p. 106). Along the western side metamorphic effects are

[1] *See*, for example, the section near Burnt Creek (Harrison, 1952, Fig. 3).

FIGURE 1.—INDEX MAP
Location of Labrador trough and Ahr Lake area

slight (Harrison, 1952, p. 17), and this condition prevails in the main portion of the trough to about the western limit of the predominantly volcanic region. Here metamorphism changes rather abruptly, possibly in part because of fault displacements, to the greenschist facies. Most of the volcanic belt is in this facies, but along the trough's eastern margin metamorphism locally grades into the epidote-amphibolite facies (Fahrig, 1957, p. 115). Both Fahrig (1957, p. 116) and Bergeron (1957b, p. 104) consider that the schists and gneisses to the east are more highly metamorphosed equivalents of the rocks of the trough.

GEOLOGY OF THE AHR LAKE AREA

General Statement

The location of the Ahr Lake area is shown in Figures 1 and 2, and the geology is indicated in Plate 1. It is an area of approximately 340 square miles, astride much of the trough's igneous belt and therefore ideally situated for a study of the igneous rocks. A more detailed map and report are available (Baragar, 1958).

General Geology

Nonintrusive rocks of the area can be grouped into three divisions: a predominantly

sedimentary group lying west of the lineament formed by Murdock and Irene lakes; a predominantly volcanic group between the Mur-

All rocks dip easterly, and, with the exception of the biotite schists, whose age relationships are unknown, the nonintrusive rocks become

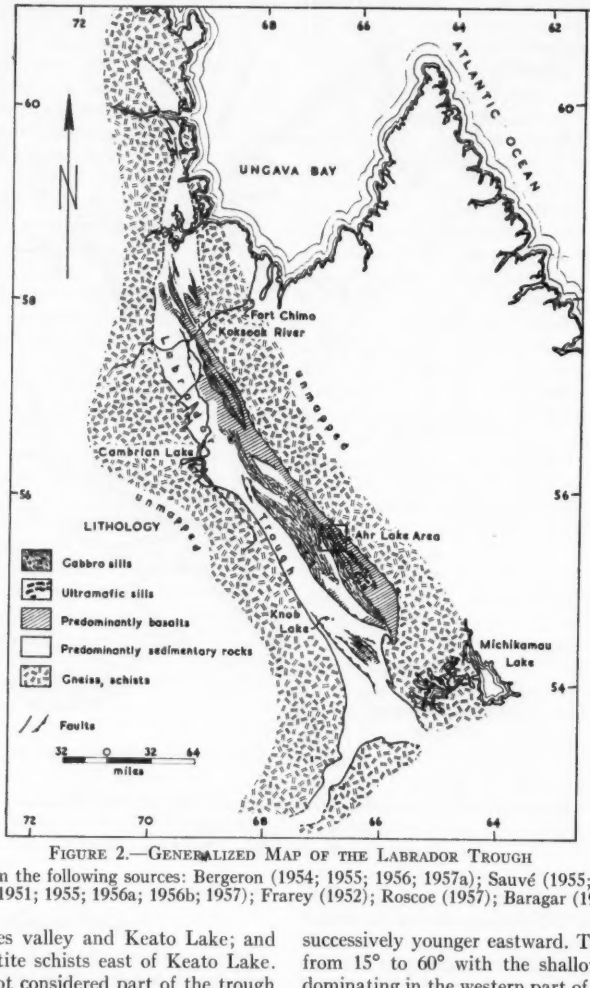

FIGURE 2.—GENERALIZED MAP OF THE LABRADOR TROUGH

Compiled from the following sources: Bergeron (1954; 1955; 1956; 1957a); Sauvé (1955; 1956); Bérard (1957); Fahrig (1951; 1955; 1956a; 1956b; 1957); Frarey (1952); Roscoe (1957); Baragar (1958).

dock-Irene lakes valley and Keato Lake; and a group of biotite schists east of Keato Lake. The latter is not considered part of the trough but belongs to the schists and gneisses which adjoin the trough on the east. The geology in the other two groups is complicated by swarms of closely spaced sills. Three belts of sills are present, each with its own particular association. The eastern belt intrudes the predominantly volcanic region and is composed of gabbro and serpentine sills; the central belt consists of gabbro and leopard-rock sills, while a western belt has gabbro sills only.

successively younger eastward. The dips range from 15° to 60° with the shallower dips predominating in the western part of the area.

The stratigraphic succession is shown in Table 1 together with the Knob Lake section. The oldest rocks, on the western side of the map area, are a succession of dark-gray to black quartz wackes and shales interlayered with basalt flows. They are overlain by a dolomite formation which can probably be correlated with the Denault formation of the type section at Knob Lake. This is followed in the Hurst Lake region by a thin shale member, then

by iron formation. The iron formation is probably equivalent to the Sokoman formation of Knob Lake. East of Lac Chassin, iron formation is missing, and dolomite is succeeded by the Howse group. The Howse group here consists of a lower member composed predominantly of wackes and shale with one intercalated bed of pyroclastic rocks and an upper member of gray basalt. The basalt is identical in appearance with that in the oldest group. The Howse group is overlain by a formation of mafic pyroclastic rocks, the Murdock formation, and this in turn is succeeded by the Doublet group, consisting of a lower sedimentary and an upper basaltic formation.

Several strike faults and a number of northwesterly and north to northeasterly oblique faults are present (Pl. 1). The trace of the faults may be marked by shallow depressions, but the faults are not exposed, and their nature and attitude is largely uncertain. The oblique faults evidently dip steeply, as they cross gabbro ridges with little deflection. The northwestern faults have left-hand displacements, the north to northeastern faults both right- and left-hand displacements. Only the strike fault northeast of Lac Harvut shows evidence of its movement. Immediately north of the gully marking the fault is a tightly folded anticline, 100–300 feet across, which parallels the fault for at least 1000 feet. The anticline is overturned to the southwest and plunges gently southeast. Assuming that this is a drag fold the eastern side of the fault must have moved up and slightly south, and the fault is probably a northeastward dipping thrust. Possibly the other strike faults are also thrusts; if so, their remarkably straight traces indicate that they are high-angle thrusts. The presence of the Murdock-Irene lakes fault is speculative. It is placed along the strike extension of the Walsh-Connolly lakes fault recognized farther south (Frarey, 1952, p. 7), but the main evidence for its existence here is the abrupt change in metamorphism found in crossing the Murdock-Irene lakes valley.

Three zones of regional metamorphism are superimposed on geology of the area. West of Murdock and Irene lakes the rocks show little apparent effect of metamorphism; between Murdock and Keato lakes they are in the green-

TABLE 1.—STRATIGRAPHY OF KNOB LAKE AND AHR LAKE AREAS*

Formation	Dominant lithology	Thickness (feet)	Formation	Dominant lithology	Thickness (feet)
			Upper Doublet	Pillowed lava, minor shales	17,000
			Lower Doublet	Graywacke and argillite	1500
			Murdock	Iron formation	0–650
				Pyroclastic rocks	6000
Menihek	Slate	1000	Howse group	Basalt, argillite, pyroclastic rocks, and graywacke	7000
Sokoman	Iron formation	550		Iron formation	0–650
Ruth	Slate	50			
Wishart	Quartzite	100			
Fleming	Chert breccia	200			
Denault	Dolomite	600		Dolomite	4200
Attikamagen	Slate	1200		Quartz wacke and shale	2000
Total thickness		3700			37,700–39,000

*Thickness in the Ahr Lake section determined by map width and dip. Unrecognized repetition by faulting is possible. Knob Lake section from Harrison (1952), thickness measured near Knob Lake town. Correlation between sections assumes iron formation and dolomite as common stratigraphic zones.

schist facies but are unfoliated; northeast of Keato Lake the rocks are well-foliated biotite schists. Gabbro sills and basalts in the greenschist facies are here called metagabbros and metabasalts.

Field Characteristics

Basalts

WESTERN ZONES: Basalts interlayered with the Howse group and with the oldest group are identical in appearance and are referred to as the western zones. They are generally light-gray, buff-weathering, dense, aphanitic rocks. With the exception of poorly defined flow contacts, meager local development of variolites, and a few pillowed layers they are massive, and amygdules appear to be absent. Locally the rock coarsens to doleritic grain size, but no evidence was found in these places that it is intrusive. The light color of the rocks suggests that they are more felsic than proved by laboratory work. Locally where the rock is dark green it is indistinguishable from the Doublet metabasalts. No pyroclastic rocks were observed with the oldest basalt member, but the Howse group basalt is underlain by a thin pyroclastic layer which interfingers with the sediments below and is overlain by the Murdock formation.

DOUBLET METABASALTS: The Upper Doublet formation is composed almost entirely of dark green, commonly pillowed, metabasalts. Coarse-grained pyroclastic beds are rare, but thin shaly members, some of which may be tuffaceous, are found at a number of stratigraphic levels. Commonly such shaly members have furnished sites for sill intrusion. Few flow contacts were seen, and amygdaloidal bands are rare and may have weathered out. Shallow linear depressions, parallel to the strike, are regular features of the topography and might represent weathered flow contacts or possibly interbedded sediments. Small, irregular bodies with gabbroic grain size are common associates of the basalts. Their origin is uncertain, but they are probably shallow intrusions contemporaneous with volcanism rather than associates of the later sills.

Basalts with sparsely distributed clots of feldspathic material, 0.2–5 cm in diameter, are found at three levels in the Upper Doublet formation. In most places 1–12 clots occur per square yard of outcrop, but locally they are absent and the rock is then similar to normal Doublet basalt. The two lower members are predominantly massive flows or possibly shallow intrusions, whereas the upper member is well pillowed. Gabbroic phases which may or may not contain feldspathic aggregates are common associates of the "blotchy" flows. In two places the gabbroic phase is spectacular. White aggregates of feldspar phenocrysts up to 8 cm in diameter are regularly spaced in a coarse-grained feldspathic gabbro (Pl. 2, fig. 1). The similarity of this phase to leopard rock is striking.

The metabasalts form an irregular rolling surface with a maximum relief of 250 feet. Metagabbro sills generally rise 50–100 feet above adjoining basalts in low rounded ridges, and ultramafic sills may form ridges or conspicuous linear depressions. Outcrop is abundant and gives the surface a knobby or bulbous appearance.

Intrusive Rocks

GENERAL PHYSICAL FEATURES: Gabbro sills form conspicuous ridges which emphasize structures in the less resistant sedimentary or vol-

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PLATE 2.—BLOTCHY METAGABBRO AND AHR LAKE SILLS

FIGURE 1.—Gabbro associated with flows of blotchy metabasalt. Northeast of Ahr Lake

FIGURE 2.—View of columnar jointing showing hexagonal pattern. West of Hurst Lake

FIGURE 3.—Looking northward along scarp face of gabbro sill. West of Lac Chassin

FIGURE 4.—Contact of a gabbro sill with underlying shale. The dip is gently east. West border of Ahr Lake area.

PLATE 3.—LEOPARD ROCKS AND CONTACT BETWEEN SUBPEGMATITE AND EVEN-GRAINED GABBROS

FIGURE 1.—Typical leopard rock. Rounded clots of altered plagioclase scattered through a coarse-grained, feldspathic groundmass. Just west of the south part of Murdock Lake in Ahr Lake area.

FIGURE 2.—Flow structure at boundary of leopard rock and gabbroic border zone. Just west of the south part of Murdock Lake in Ahr Lake area.

FIGURE 3.—Contact between subpegmatite gabbro (left) and even-grained gabbro (right). Specimen R67. Plain light. ×4.

FIGURE 2

FIGURE 4

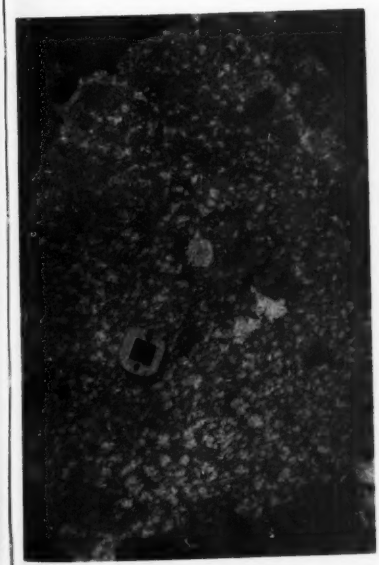


FIGURE 1

FIGURE 3

BLOTCHY METAGABBRO AND AHR LAKE SILLS

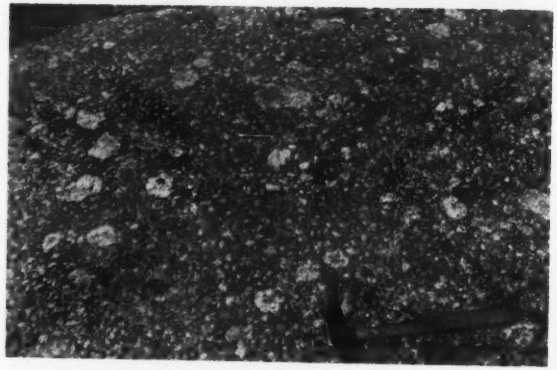


FIGURE 1

FIGURE 2

FIGURE 3

LEOPARD ROCKS AND CONTACT BETWEEN SUBPEGMATITE AND EVEN-GRAINED GABBROS

FIGURE 1

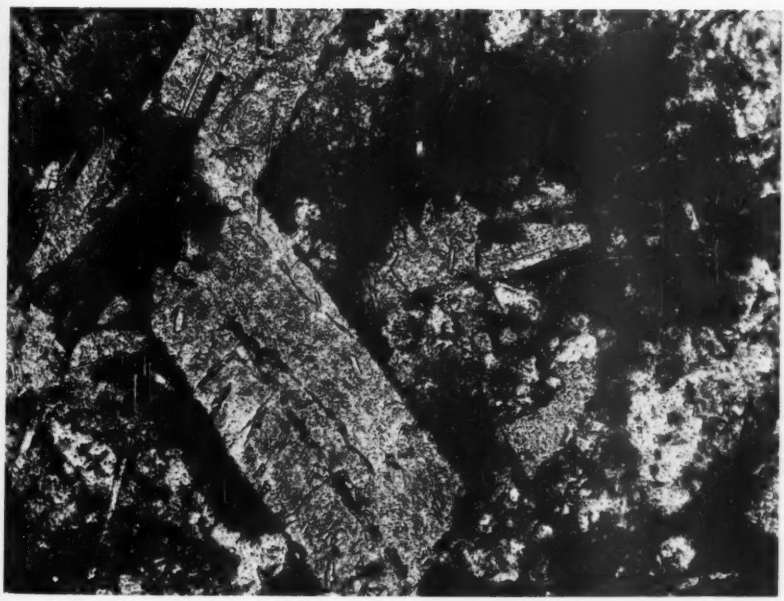


FIGURE 2

THIN SECTIONS OF PEGMATITIC GABBROS

FIGURE 2

LEOPARD ROCKS AND CONTACT BETWEEN SUBPEGMATITE AND EVEN-GRAINED GABBROS

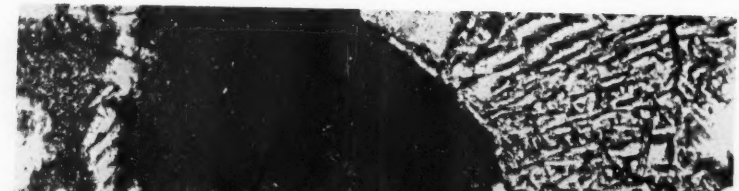


FIGURE 1

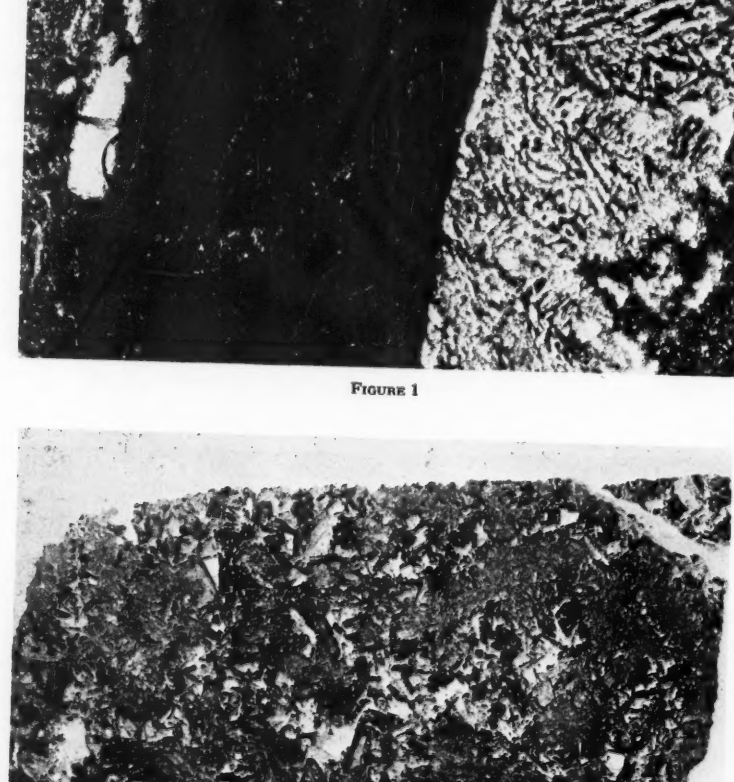


FIGURE 2

ALBITE GABBRO PEGMATITE AND SEGREGATIONS IN GABBRO

canic host rocks. Dipping east to northeast the sills present southwest-facing escarpments ribbed with columnar joints (Pl. 2, figs. 2, 3, 4). Many planar joints are at right angles to the columnar jointing and may present an appearance of stratification. Sedimentary interlayers weather more readily than adjoining gabbros and give rise to elongate, lake-filled valleys, which characterize the topography of the eastern half of the Labrador trough. In the Ahr Lake area ridges in the western belts are 100–400 feet high, and along the western border of the area where dips are low (15°–30°) they form distinct cuestas. Metagabbros of the eastern belt form rounded hogbacks rarely exceeding 100 feet in height.

The thickness of sills ranges from a few feet to about 1500 feet. Elsewhere in the Labrador trough maximum thicknesses of 1000 feet (Kavanagh, 1954, Ph.D. thesis, Princeton Univ., p. 30 a) and 3000 feet (Sauvé, 1955, p. 4) have been recorded. Sauvé (1955, p. 4) suggests that the thicker sills are multiple sills in which each intrusion is only 300–500 feet thick.

TYPES OF GABBRO: The gabbros can be broadly divided on the basis of field characteristics into normal gabbros, leopard rocks, and metagabbros.

NORMAL GABBROS:[2] Normal gabbros are generally fairly fresh, gray-green, medium-grained rocks found in the western and central belts. They may range within single sills from sparsely prophyritic and ophitic rock types near the base to pegmatitic phases in the upper portions. Three to 10 feet of chilled marginal phase has been observed at both upper and lower contacts of several sills. Calcite amygdules were found in the upper chilled margin at two or three localities and argillite xenoliths at a few others. The xenoliths show little reaction with the gabbro.

LEOPARD ROCKS: Leopard rocks are coarse-grained feldspathic gabbros regularly spotted with roughly spherical aggregates of altered feldspar 6–15 cm across (Pl. 3, fig. 1); 1–6 aggregates occur per square yard of outcrop. Many sills of leopard rock contain a core zone of normal gabbro. In the first sill east of the Hurst Lake-Lac Chassin valley (Pl. 1) a sinuous layer of gabbro, flanked on both sides by leopard rock, forms the center of the sill. It thins and thickens and swings from side to side along its length. Northwest of Youngren Lake it lenses out, then reappears farther along. In detail it interfingers with leopard rock in some places (Fig. 3). Three other sills in the central belt of gabbros seem to have a similar structure, although their relationships are not so clear owing to crowding of sills and uncertainty of the positions of their boundaries. Just west

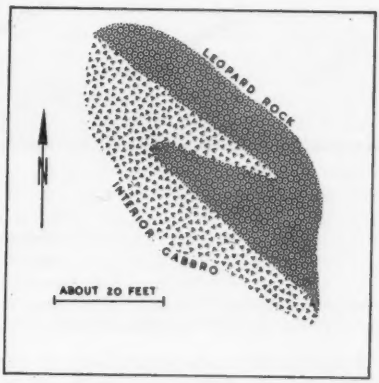


FIGURE 3.—INTERFINGERING OF LEOPARD ROCK AND INTERIOR GABBRO

From a field sketch of the upper contact directly east of Lac Chassin.

[2] The term "normal gabbro" applies to the Ahr Lake area only where it designates the most common variety of gabbro. Classification of these rocks as gabbros rather than dolerites is based on grain size (Hatch, Wells, and Wells, 1949, p. 212, 294).

PLATE 4.—THIN SECTIONS OF PEGMATITIC GABBROS

FIGURE 1.—Gabbro pegmatite (R554) with micropegmatitic intergrowths radiating from plagioclase (lower center). Chlorite and pyroxene are not well defined. Crossed nicols. ×20.

FIGURE 2.—Albite-gabbro pegmatite (R83) showing dark mesostasis and slightly clouded albite. Apatite, quartz, and chlorite are other minerals. Plain light. ×110.

PLATE 5.—ALBITE GABBRO PEGMATITE AND SEGREGATIONS IN GABBRO

FIGURE 1.—Albite-gabbro pegmatite (R11) showing finely textured micro-pegmatite radiating from albite. Crossed nicols. ×75.

FIGURE 2.—Even-grained gabbro (R130) with small pockets of late-crystallizing minerals which appear as lighter colored areas. Plain light. ×4.

of Murdock Lake what seems to be a single sill has a thin layer of leopard rock at its upper margin, a central zone of gabbro, and a wide lower margin of leopard rock. Other examples are south of Lac Villeneuve and on the east side of Swanty Lake.

METAGABBROS: Metagabbros in the eastern group of sills are characterized by a finely spotted or speckled appearance because of color contrast between altered feldspar and green amphibole. Commonly amphibole forms discrete, rounded spots set in a blue-gray

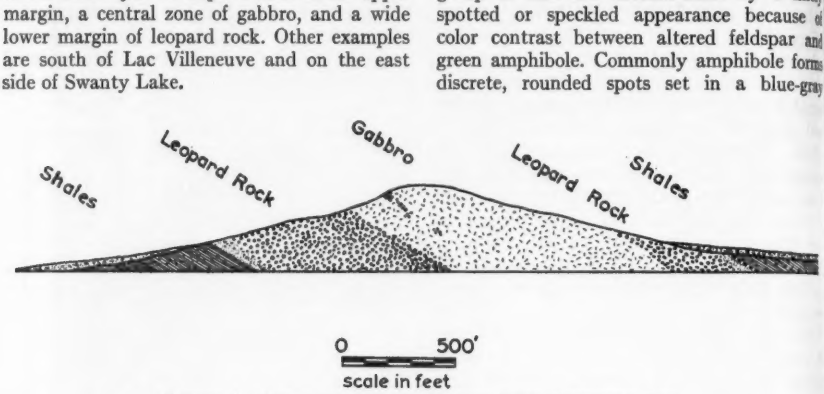


FIGURE 4.—CROSS SECTION OF LEOPARD-ROCK COMPOSITE SILL

Gabbro has intruded the center of a leopard-rock sill and locally isolated lenses of leopard rock. Thin zones of gabbro are at the margins of the sill. Northeast of Hurst Lake.

Chilled borders of leopard-rock sills are dark basalts without visible phenocrysts. Within 2–10 feet of the contact the rock grades into dark, medium-grained gabbro with sparsely distributed feldspar phenocrysts and aggregates of phenocrysts. In a marginal zone 30–75 feet wide the rock is predominantly gabbro, but inward from this the number and size of phenocrysts and aggregates increase sharply until they coalesce to form leopard rock. In at least one place leopard rock grades into marginal gabbro through a zone showing a laminar arrangement of feldspar phenocrysts (Pl. 3, fig. 2). As leopard rock passes into this zone the feldspar aggregates appear to be broken down into individual crystals which are strung out into streaky layers parallel to the sill margins. This is interpreted as flow structure.

Contacts of leopard rock with the interior gabbro are abrupt but not sharp. Leopard rock passes into gabbro at a surface marking the limit of coarse feldspathic aggregates. Within the gabbro close to the contact are scattered clots and large feldspars which have apparently been wrenched from the adjoining leopard rock. Isolated lenses of leopard rock roughly parallel to the sill margins may be encountered within the interior gabbro; one lense is cut by a vaguely defined gabbro dike which seems to be continuous with the surrounding gabbro.

The leopard rock-gabbro sills are interpreted as composite sills; leopard rock is intruded by later normal gabbro. Figure 4 gives a cross section of the most westerly composite sill.

groundmass, in a polka-dot pattern. Elsewhere amphibole forms irregular interconnecting patches. Foliated metagabbros are rare. Near contacts the rock is finer-grained but not aphanitic.

Two distinctive types of metagabbro are termed melanometagabbro and leucometagabbro. Typical melanometagabbro consists of large interlocking crystals, 0.5–3 cm in diameter, of grayish-green amphibole. Few other minerals can be seen in hand specimen. Leucometagabbro in its most spectacular development is light cream to white, sparsely dotted with greenish blebs. In other places it is bluish gray and inconspicuously marked with slightly darker amphibole crystals. Melanometagabbro occupies the lower portion, and leucometagabbro the upper portion, of two separate sills; one a mile west of Valley Ponds, the other on the west side of Protecto Lake (Pl. 1). In the Valley Ponds sill the upper member extends several miles beyond the lower member.

Melanometagabbro grades up into leucometagabbro in the Protecto Lake sill, and although the contact was not seen in the Valley Ponds sill, presumably the same relationship exists there. The major part of the gradation takes place across a zone about 200 feet wide; the sill itself is more than 1500 feet thick (assuming a dip of 50°, the average in adjoining pillow lavas). Thus the sill is essentially composed of two fairly uniform rock types. Similarly, each of the two zones in the Valley Ponds sill is composed of a uniform rock type.

Each variety of metagabbro may form

independent sills. Typical melanometagabbro is found without associated leucometagabbro about a mile west of the Valley Ponds sill. The boundaries are complicated by faulting; hence the sill was not necessarily emplaced separately. A leucocratic metagabbro, similar to that in the upper portions of the Valley Ponds and Protecto Lake sills forms much of the sill just east of Lac Harvut but has no corresponding melanocratic base.

Petrography

Basalts

WESTERN ZONES: The light-gray basalts of the western zones present a distinctive appearance in thin section. An extremely fine-grained groundmass is crowded with plagioclase microlites 0.2–0.5 mm long and commonly hollow. Cross sections are hollow squares 0.02–0.04 mm to a side, filled with dark material similar to the groundmass. Rare, rounded to polygonal serpentine patches are probably olivine pseudomorphs.

In most places the groundmass is a fuzzy brownish mass, but in a few sections the matrix is clear and shows slender curving pyroxene rods in fanlike clusters; in places single pyroxene rods give rise to whole families of fibers which sweep out into graceful curving sprays. Minute plagioclase needles are thickly packed on all sides with parallel rods of pyroxene which extend at right angles from the plagioclase surface. The plagioclase crystallites may have acted as nuclei for pyroxene crystallization. The groundmass is charged with tiny specks of brownish opaque material, probably leucoxene-magnetite. The textures can be matched by Battey's (1956, p. 90) descriptions of New Zealand spilites.

Coarser-grained subophitic phases of the lavas bear little resemblance to the aphanites. Pyroxene forms equidimensional crystals between plagioclase laths. Triangular intergrowths of magnetite and sphene are sparsely scattered throughout. Rare blebs of serpentine may be found, and some show meshwork structure indicating derivation from olivine.

Alteration is inconspicuous. Plagioclase may be partly replaced by dusty, brownish epidote, less commonly by serpentine and chlorite. Pyroxene may be streaked and patchy with chlorite replacement. Rare serpentine is found at the center of pyroxene crystals and is interpreted as pseudomorphs of pigeonite.

Chemical and spectrographic analyses of the coarser-grained lava are presented in Tables 8 and 11 (B54A).

DOUBLET METABASALTS: Lavas of the Doublet group are composed of albite, actinolite, epidote, chlorite, and finely granular sphene or leucoxene. Surviving igneous textures are not common. Most thin sections show a randomly oriented mesh of shredded actinolite blades with dusty granular mixtures of epidote and albite and patches of chlorite. A few sections show vaguely defined, lathlike pseudomorphs after plagioclase. One thin section of Doublet metabasalt (loaned by W. F. Fahrig) from the Griffis Lake area is less altered than most and still contains relics of hollow and forked plagioclase crystals similar to those in lavas of the western zones. Feldspathic clots of "blotchy" metabasalt, originally clusters of plagioclase phenocrysts, are replaced by a dusty mixture of epidote and albite. The groundmass is similar to the even-grained lavas.

Normal Gabbros

Several types of gabbro encountered can be attributed to differentiation *in situ*. Similar varieties were found at roughly corresponding levels in different sills. The following account of the differentiation is based on observations and specimens from many sills but presumably could have been furnished from an ideal section of a single sill (Fig. 5). Portions of this section have been observed in a number of sills, and the range given for each division is an average of observed ranges. Descriptions of differentiation products are given below.

PORPHYRITIC GABBRO: This variety is found at the base of a number of sills. It is a dark rock with sparse plagioclase phenocrysts, 2–6 mm across, and ophitic pyroxenes up to 3 cm in diameter. In most hand specimens 2–12 phenocrysts are visible. In many places the rock is spotted with serpentine blebs derived from olivine. Chemical analyses and norms are given in Table 11 and volumetric analyses in Table 2.

Olivine ranges from 4 to 18 per cent, combined pyroxenes from 20 to 30 per cent, and plagioclase from 50 to 60 per cent. Iron ores are generally minor. Plagioclase phenocrysts constitute an estimated 1–10 per cent of the rocks.

A typical porphyritic gabbro is shown in Figure 6A. Many orthopyroxene crystals contain oriented exsolution streaks and blebs of augite which may persist even when the crystal has been replaced by serpentine. Augite with exsolved orthopyroxene has not been observed. Iron ores are generally brownish masses of

sphene containing blebs or triangularly arranged lamellae of ilmenite. Some masses reach 1 mm in diameter and are ophitic or subophitic.

In thin sections the rock contains large, equidimensional augites packed with randomly oriented plagioclase laths and a few serpentine patches after olivine. The large augite crystals

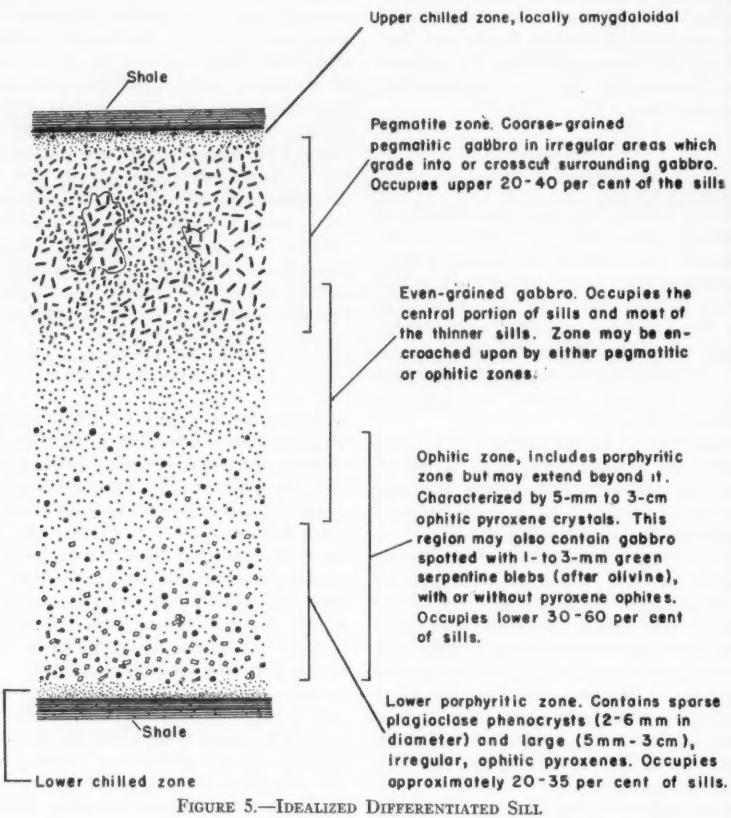

Upper chilled zone, locally amygdaloidal
Shale
Pegmatite zone. Coarse-grained pegmatitic gabbro in irregular areas which grade into or crosscut surrounding gabbro. Occupies upper 20-40 per cent of the sills
Even-grained gabbro. Occupies the central portion of sills and most of the thinner sills. Zone may be encroached upon by either pegmatitic or ophitic zones.
Ophitic zone, includes porphyritic zone but may extend beyond it. Characterized by 5-mm to 3-cm ophitic pyroxene crystals. This region may also contain gabbro spotted with 1- to 3-mm green serpentine blebs (after olivine), with or without pyroxene ophites. Occupies lower 30-60 per cent of sills.
Lower porphyritic zone. Contains sparse plagioclase phenocrysts (2-6 mm in diameter) and large (5 mm-3 cm), irregular, ophitic pyroxenes. Occupies approximately 20-35 per cent of sills.
Shale
Lower chilled zone

FIGURE 5.—IDEALIZED DIFFERENTIATED SILL

OPHITIC GABBRO: This rock type extends up to the central portion of many sills and may include the porphyritic member at the base (Fig. 5). In outcrop the rock has an irregular surface owing to differential weathering of plagioclase and ophitic pyroxenes. The fresh surface is mottled grayish green with the characteristic ophites inconspicuous unless cleavages reflect the light. Several ophitic pyroxenes up to 3 cm in diameter occur in some hand specimens. Olivine, or its serpentine pseudomorph, is present in most places but not an essential constituent. In a few places the rocks are thickly spotted with brownish-green serpentine blebs and are then called spotted gabbro.

occur singly and are separated by areas of even-grained, subophitic gabbro. A second pyroxene is rare, but evidence of its former existence in the form of serpentine patches, some with oriented augite lamellae, is not uncommon. Serpentine patches which are pseudomorphs after olivine have a characteristic rounded shape and meshwork structure; less regular patches are assumed to represent lime-poor pyroxene. Opaque minerals, principally ilmenite-sphene intergrowths, are again scarce. Figure 6B shows a typical thin section.

Volumetric analyses of ophitic and spotted gabbro specimens are given in Table 2 and chemical analyses and norms in Table 11. Specimens B295 and R99 are typical spotted

gabbros from positions low in their respective sills. Mineral contents are not too different from those of the porphyritic variety.

EVEN-GRAINED GABBRO: Even-grained gabbro is the most common variety and is found in the central and upper parts of sills. The rock consists almost entirely of subophitic pyroxene and plagioclase of similar grain size. Pyroxene is predominantly augite, but orthopyroxene relics are found in two thin sections. Areas of serpentine, some containing oriented augite lamellae, are present in most thin sections and presumably represent orthopyroxene or pigeonite. Twins are still preserved by chevron-patterned strings of augite blebs or lamellae in serpentine. Such pyroxene was probably pigeonite which commonly twins on (100). Intergrowths of ilmenite and sphene are slightly more abundant in this rock type. Volumetric data of even-grained gabbros are given in Table 2, and Figure 6C illustrates the rock type.

SUBPEGMATITIC GABBRO AND GABBRO PEGMATITE: The upper 20–40 per cent of many sills contains abundant pegmatitic gabbro in irregular isolated bodies or interconnecting masses a few inches to many feet across. Pegmatite grades into or crosscuts even-grained gabbro. Crosscutting boundaries are abrupt but not knife-edged; individual crystals are continuous across the boundary but change character in passing from one rock to the other (Pl. 2, fig. 3). Figure 5 illustrates the patchy character of pegmatite distribution.

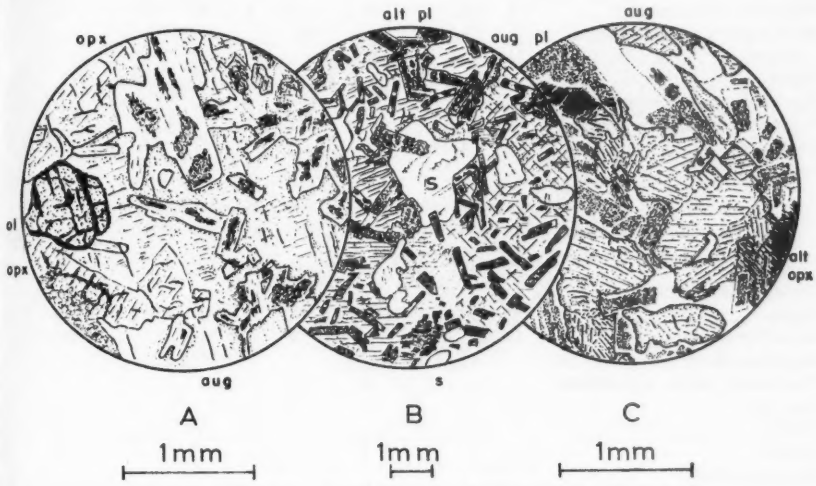


FIGURE 6.—PORPHYRITIC, OPHITIC, AND EVEN-GRAINED GABBROS

A. Porphyritic gabbro showing olivine and plagioclase embedded in augite. Specimen B184. B. Ophitic gabbro spotted with pseudomorphs after olivine. Shows large pyroxene enclosing altered plagioclase laths and serpentine blebs after olivine. Specimen B295. C. Even-grained gabbro showing subophitic augite and plagioclase: Altered patches may have been orthopyroxene. Specimen R90A.
ol: olivine; opx: orthopyroxene; aug: augite; pl: plagioclase; alt: altered; s: serpentine

The pegmatites are called gabbro pegmatite or subpegmatitic gabbro depending on the presence or absence of micropegmatite. Otherwise they are similar and resemble gabbro pegmatites described from other intrusive rocks (Walker, 1953; Tomkeieff, 1929, p. 102–109; Broderick, 1935, p. 510–511; Cornwall, 1951, p. 163). Subpegmatitic gabbro is the more abundant and may occupy most of the upper part of sills. Gabbro pegmatite seems to occur in relatively small pockets within subpegmatitic gabbro.

The pegmatites are characterized by elongated (2–10 mm) bent and twisted pyroxene, generally twinned along the medial line of the elongated crystal so that (001) striations form a herringbone pattern. The rock has a high iron content and is dark, in contrast with more mafic varieties of gabbro which are generally light greenish gray.

Figure 7(A, C) shows typical thin sections of subpegmatitic gabbro and gabbro pegmatite. Elongated, centrally twinned augite is dominant and forms a rough framework for other

TABLE 2.—VOLUMETRIC ANALYSES OF NORMAL GABBROS

	Porphyritic Gabbro				Ophitic Gabbro					
	B343	B184	B279	B154	B345	R90C	B295	R99	R126	B111
Plagioclase	51	53	62	49	56	46	49	63	50	60
Olivine (+ pseudomorphs)	4	13	15	18	4	3	13	13	2	..
Orthopyroxene (+ pseudomorphs)	34	11	11	26	..	..	..	..	..	..
Augite		18	9		22	39	29	19	24	27
Pigeonite (+ pseudomorphs)						3				
Serpentine (uncertain origin)	5	..	..	..	15	5	8	2	19	6
Hornblende, biotite, and chlorite	5	4	2	5	2	2	..	..	3	4
Iron ores	1	1	1	2	1	2	1	3	2	3

	Even-Grained Gabbro				Subpegmatitic Gabbro				Gabbro Pegmatite	
	R34	R64	R100	R153	R286	R23	R129	B152*	R554	B133
Plagioclase	51	48	63	58	46	41	41	53	35	39
Olivine pseudomorphs	..	..	2	..	..	..	..	..	..	..
Augite	33	33	27	26	28	39	21	26	18	27
Serpentine (after pigeonite?)	13	..	..	7	7	..	11	..	..	..
Serpentine (uncertain origin)	..	17	5	..	..	..	..	..	..	..
Hornblende, biotite, and chlorite	..	..	..	6	..	..	..	..	..	..
Iron ores	3	2	3	2	4	5	5	5	4	8
Chlorite	..	..	..	..	8	9	10	15	15	13
Quartz	..	..	..	1	7	6	12	1	28	13
Micropegmatite									(32)	(9)

	Albite gabbro pegmatite		
	R93†	R151	R94
Plagioclase	44	41	34
Augite	28	25	10
Iron ores	11	10	11
Chlorite	15	18	20
Quartz	2	2	4
Apatite	n.d.	n.d.	2
Pumpellyite	..	4	..
Mesostasis	..	..	19

	Averages					
	Porphyritic gabbro	Ophitic gabbro	Even-grained gabbro	Subpegmatitic gabbro	Gabbro pegmatite	Albite gabbro pegmatite
Plagioclase	54	55	55	44	38	38
Pyroxene**	33	37	41	34	22	18
Olivine**	12	6	0.5	..	..	..
Iron ores	1	2	3	5	6	11
Quartz	..	..	0.5	8	20	3
Chlorite	..	..	..	9	14	19
Mesostasis	..	..	..	..	..	10
Apatite	..	..	..	..	..	1

* Transitional between even-grained and subpegmatitic gabbros

† Transitional between even-grained gabbro and albite gabbro pegmatite

** Pyroxene includes serpentine (other than that pseudomorphous after olivine) and other mafic accessories. Olivine comprises olivine and olivine pseudomorphs.

minerals. Plagioclase is elongated but less conspicuous than pyroxene. Interstices of subpegmatitic gabbro are filled with wedges and rounded blebs of quartz, rare micropegmatite, but the rocks are altered. Augite is close to ferroaugite in composition; hence possibly only one pyroxene crystallized.

Volumetric data are given in Table 2, and

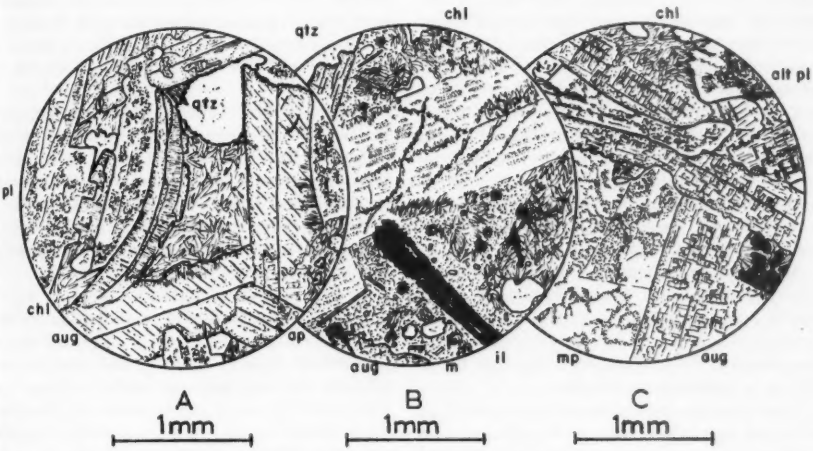


FIGURE 7.—PEGMATITIC GABBROS

A. Subpegmatitic gabbro. Shows the dominance of elongated, twinned, and bent augites; plagioclase is less conspicuous. Quartz and high birefringent chlorite fill interstices. Specimen B286. B. Albite gabbro pegmatite. Shows a slightly clouded albite crystal veined and party replaced by chlorite. The surrounding interstitial space is filled with high birefringent chlorite, unidentified mesostasis, apatite, quartz, and iron oxide. Platy ilmenite in the lower portion of the field is typical of iron ores in albite gabbro pegmatite. Specimen R94. C. Gabbro pegmatite. Elongated and twinned augite crystals (partly altered) are the dominant feature. Micropegmatite fills interstices on the left side of the field, chlorite on the upper, and chlorite and iron ores on the lower right side. Plagioclase is andesine. Specimen R554.

pl: plagioclase; qtz: quartz; chl: chlorite; aug: augite; mp: micropegmatite; il: ilmenite; m: mesostasis; ap: apatite; alt: altered

and a peculiar, high-birefringent, iron-rich chlorite (Table 5). Embedded in chlorite or quartz are numerous tiny apatite rods with pleochroic halos. Apatite was not observed in more mafic varieties of gabbro. Altered iron ores are abundant and form large (up to 3 mm), interstitial masses of sphene with globules or triangularly arranged lamellae of ilmenite. Gabbro pegmatite contains, in addition, abundant micropegmatite (Pl. 4, fig. 1), and the iron-rich chlorite shows deeper colors and more extreme pleochroism.

Augite is the only pyroxene in subpegmatitic gabbro, but enclosed serpentine (or chlorite) patches may represent altered pigeonite. Some of the patches show feebly defined, curving, and branching lines reminiscent of fractures in pigeonite. Walker (1953, p. 46) finds that pigeonite is the normal lime-poor pyroxene of gabbro pegmatites and commonly forms altered cores in augite. No evidence of lime-poor pyroxene was observed in gabbro pegmatite, chemical analyses and norms are listed in Table 11. Specimen B152 (Table 2) has the texture of even-grained gabbro but contains iron-rich chlorite and has abundant iron ores; hence it is probably transitional to subpegmatitic gabbro.

ALBITE GABBRO PEGMATITE: This variety of gabbro is characterized by sodic plagioclase. All occurrences in the Ahr Lake area belong to the upper portion of a folded sill north-northwest of Hurst Lake. The main portion of the sill exposed in the west-facing scarp northwest of Lac Bourlamaque is even-grained gabbro similar to that in other sills. On top of the escarpment and along the gently dipping east slope, rather dark albite gabbro pegmatite can be found in most outcrops up to the point where the sill disappears below dolomite just north of Lac Bourlamaque. It outcrops again in the anticlinal nose north of Hurst Lake and can be found along the east limb as far north as Float Bay.

The even-grained gabbro has a uniform

appearance throughout the 150 feet exposed below the albite gabbro pegmatite layer. Neither porphyritic nor ophitic gabbros were observed in the lower part of the scarp face, but the base of the sill is not exposed. The gabbro is severely altered. Plagioclase is extensively saussuritized and locally replaced by minor pumpellyite and prehnite. Augite is commonly discolored by a brownish lamellar alteration and is partly replaced by chlorite or amphibole. Serpentine, chlorite, prehnite, and unidentified material form irregular patches in the rock. Even-grained gabbro apparently grades rapidly into albite gabbro pegmatite. The complete gradation was not observed, but a transitional phase was found just below albite gabbro pegmatite.

Albite gabbro pegmatite is black or dark brown and contains conspicuously elongated (up to 2 cm), randomly oriented, white, light-brown, or colorless plagioclase crystals. In a few places hexagonal plates of ilmenite up to 0.5 cm in diameter are visible.

Thin sections are dominated by stout albite laths and long, thin plates of ilmenite, set in a heterogeneous mixture of pyroxene, altered iron ores, chlorite, and other fine-grained minerals (Fig. 7B). Pyroxene is less conspicuous than in subpegmatitic gabbro and gabbro pegmatite and is extensively chloritized. Most of the albite is slightly clouded, and some is partly veined by chlorite (Fig. 7B; Pl., 4 fig. 2). Locally it shows minor alteration to pumpellyite and prehnite. Pumpellyite also occurs in interstices and is locally an important minor constituent of the rock. Chlorite with subordinate quartz is a common interstitial filling. Chlorite is partly the high-birefringent variety of the other pegmatitic phases and may show extreme pleochroism from brilliant orange brown to green (Table 5). Apatite is ubiquitous and embedded in chlorite and quartz; most has pleochroic halos. In addition to platy ilmenite, some iron ores occur as altered masses of sphene with triangularly arranged lamellae of ilmenite. A mesostasis of fine-grained chlorite, unidentified colorless material of low birefringence, and minute amphibole needles is present in many specimens (Fig. 7B). It is heavily charged with dusty or fibrous particles which may form fuzzy, radiating fringes at plagioclase margins (Pl. 4, fig. 2). Commonly the mesostasis contains irregular quartz crystals and abundant rods of apatite. It is tentatively interpreted as divitrified glass. Although most albite gabbro pegmatite contains subordinate quartz, one outcrop on the nose of the anticline directly north of Hurst Lake furnished specimens with abundant micropegmatite. This has a delicate, lacelike texture which surrounds and fans out from central plagioclase crystals (Pl. 5, fig. 1). Volumetric analyses of albite gabbro pegmatite are given in Table 2 and chemical analyses in Table 11.

Specimen R93 (Tables 2, 11) is transitional between even-grained gabbro and albite gabbro pegmatite and shows features of both. It is texturally similar to even-grained gabbro and the plagioclase is An_{50}, but it contains abundant iron ores, some of them platy, and has interstitial quartz, high-birefringent chlorite, apatite and dusty mesostasis.

VEINS AND SEGREGATIONS: Veins and segregations in the sills provide information on the late magmatic fluids. Figure 2 of Plate 5 shows pockets of quartz, chlorite, and apatite surrounded by even-grained gabbro. These are interpreted as small segregations of late magmatic fluids trapped in a framework of earlier minerals. This type of association is rare. Irregular, small veinlets observed in several thin sections similarly contain chlorite, quartz, and apatite. Most of the veinlets are discontinuous and without well-defined walls. The contents of veinlets and segregations are approximately the same as the interstitial material of the pegmatitic gabbros. The large pegmatitic bodies in the upper portion of sills may be the regions of final accumulation of the late magmatic fluids.

A 2-inch vein in the sill containing albite gabbro pegmatite weathers a light color and cuts even-grained gabbro some distance below the albite gabbro pegmatite. The vein material is a dark, fine-grained rock with sparse plagioclase laths similar to those in albite gabbro pegmatite. The thin section shows a fine-grained mixture of abundant pumpellyite, prehnite (?) iron ores, and quartz. Plagioclase is corroded and largely replaced by pumpellyite and finely divided prehnite (?). Probably the vein material is one of the latest magmatic fractions and may have solidified largely as glass.

CHILLED PHASES: The chilled phases have a small percentage of microphenocrysts set in an extremely fine-grained matrix. Plagioclase, augite, and serpentine pseudomorphs after olivine are normally present as microphenocrysts.

crysts. The volumetric analysis of a typical specimen is:

Groundmass	83
Plagioclase	12
Augite	1
Serpentine (after olivine?)	4

All specimens of chilled contacts appear to have been metasomatized. Commonly they are permeated with fine carbonate minerals, and plagioclase is generally albitized. Two samples were analyzed, but both proved to be considerably altered.

ALTERATION: The normal gabbros have all been altered to some extent. Alteration is least in the porphyritic gabbros and increases through the series to the pegmatitic phases. Fresh olivine is found in porphyritic gabbro only. Lime-poor pyroxene is commonly fresh in porphyritic gabbro and rarely fresh in ophitic, spotted, or even-grained gabbros. It was not observed in subpegmatitic gabbro, although its pseudomorphs persist throughout. Plagioclase is generally fresh in porphyritic gabbros, shows considerable alteration in the intermediate gabbros, and is rarely fresh in the pegmatitic phases (except albite in albite gabbro pegmatite).

Alteration effects on the various minerals are: Plagioclase is typically saussuritized to a grayish-brown mass, too fine-grained for precise identification. Locally pumpellyite and prehnite are important alteration products, particularly in pegmatitic gabbros. Plagioclase may be veined and partly replaced by chlorite or serpentine. Olivine is generally replaced by serpentine, less commonly by quartz, bowlingite or talc. Wilshire (1958) found that the alteration products of olivine and orthopyroxene identified variously as iddingsite, bowlingite, and serpentine are all composed of mixed-layer smectite-chlorite with varying but minor amounts of antigorite, chrysotile, goethite, magnetite, quartz, calcite, talc, and mica. The terms serpentine and bowlingite are used in this paper for alteration products which resemble these minerals; they may be smectite-chlorite as contended by Wilshire. Serpentine is a colorless, low-birefringent substance with locally anomalous blue interference colors and commonly showing meshwork structure in olivine pseudomorphs. Bowlingite is a near-uniaxial negative mineral with micaceous cleavage, yellowish-brown (X) to green (Y and Z) pleochroism, and birefringence of 0.030–0.035. Normally the original shape and structure of olivine are faithfully reproduced by the replacing mineral. Even quartz is a pseudomorph after olivine. Orthopyroxene and pigeonite are normally replaced by serpentine and less commonly by chlorite and talc. Chlorite and serpentine are probably both smectite-chlorite (Wilshire, 1958), but the terms as used here apply to green-pleochroic and colorless varieties respectively. Chlorite is the more common replacement of lime-poor pyroxene in subpegmatitic gabbro, whereas serpentine is more common in mafic gabbros. Augite is little altered in most specimens. Striations due to alteration along (001) and brownish discoloration around crystal margins are usually the only effects. In gabbro pegmatite or albite gabbro pegmatite alteration is more severe, and augite may be partly replaced by chlorite. Amphibole is a rare alteration product. Poldervaart and Hess (1951, p. 481) identified (001) striations in augite as pigeonite (or inverted pigeonite) exsolution lamellae. Exsolved pigeonite lamellae have not been seen in Ahr Lake augites, but lamellar alteration on (001) may be their equivalent. Powder photographs were made of augites with well-developed striations, and the only lines obtained besides pyroxene could be assigned to chlorite. Probably the lamellar alteration results from replacement of (001) pigeonite lamellae by chlorite or serpentine. Iron ores are extensively altered to a murky, brownish, translucent substance, most of which is finely granular sphene. In this paper sphene is included in the term iron ores.

ORDER OF CRYSTALLIZATION: The chilled contacts show that olivine, pyroxene, and plagioclase were present when the magma was at least 80 per cent liquid. This agrees with the recent work of Yoder and Tilley (1957, p. 157–158), who found that the major mineral groups in mafic magmas crystallize within a narrow temperature range. Textures in the chilled phases do not indicate which of the three crystallized first; the writer assumes that the order of crystallization is represented by the change of mineralogy upward in the sill.

Olivine is confined to the early stages of crystallization where it is accompanied by both orthopyroxene and augite. Locally pyroxene surrounds olivine as a reaction rim. Orthopyroxene (largely inverted pigeonite) and augite, nearly equally abundant in the earliest rocks, probably crystallized together in equal

quantities at first. Shortly after, augite crystallization became dominant, as indicated by the sharp decline in quantities of pigeonite in the rocks later than ophitic gabbro. After the subpegmatitic stage only augite formed.

contains sparse plagioclase phenocrysts and substantial amounts of olivine (±15 per cent). Orthopyroxene with exsolved augite lamellae and augite are both present. Ores are altered titaniferous magnetite. (2) Ophitic and spotted

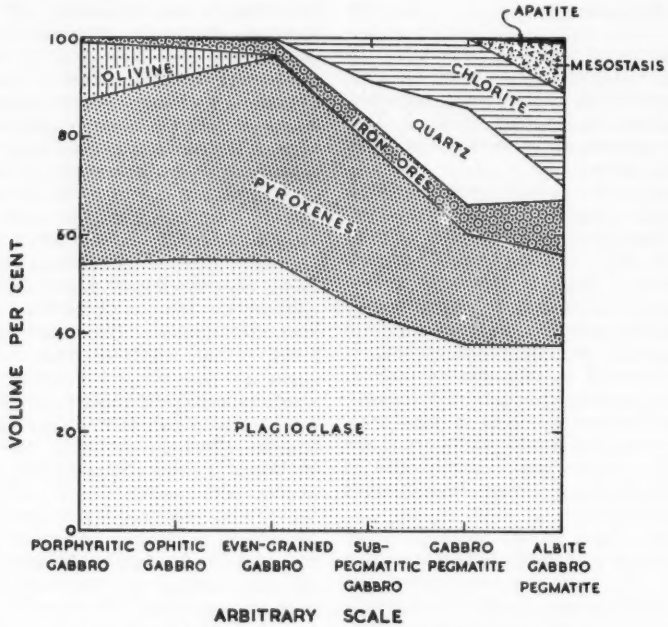


FIGURE 8.—MINERAL VARIATION IN NORMAL GABBRO SERIES
Cumulative diagram based on average volumetric analyses of each type of gabbro (Table 2)

Plagioclase and pyroxene formed together throughout the crystallization period as indicated by their ophitic or subophitic relationships. Iron ores at each stage of differentiation crystallized with the major silicates but may have finished later. They become more abundant in the late stages of crystallization.

Quartz, iron-rich chlorite, and apatite occupy interstices of the pegmatitic gabbros and are the last primary minerals to crystallize. Serpentinization of olivine and lime-poor pyroxene, and saussuritization of plagioclase may be partly due to deuteric alteration.

SUMMARY: Many of the Ahr Lake sills are differentiated. Olivine-bearing rocks occur in the lower portions of sills, pegmatitic gabbros in the upper portions. Similar products of differentiation are found at roughly corresponding levels in different sills. Salient features are: (1) Porphyritic gabbro at the base of sills gabbros occupy the lower 30–40 per cent of sills and may include the porphyritic basal zone. Ophitic gabbro is characterized by scattered ophitic augites and spotted gabbro by abundant serpentinized olivine. Subordinate lime-poor pyroxene is represented largely by serpentine pseudomorphs. Ores are in minor amounts. (3) Even-grained gabbro is a grayish-green rock occupying central portions of differentiated sills and most of the undifferentiated sills and consists of subophitic augite and plagioclase. Lime-poor pyroxene is in minor amounts, rarely as relic crystals, more commonly as pseudomorphs of serpentine. Olivine and quartz are rare. Altered titaniferous magnetite is slightly more abundant. (4) Subpegmatitic gabbro forms irregular bodies in the upper portion of sills. It is a dark rock with elongated, bent, and twisted augites. Serpentine or chlorite within augite may be after pigeonite. Interstices contain quartz, apatite, and a high-

birefringent chlorite. Titaniferous magnetite is more abundant than in rocks described above. (5) Gabbro pegmatite is in pockets in the subpegmatitic gabbro. It differs from subpegmatitic gabbro in containing micropegmatite. (6) Albite gabbro pegmatite is a dark rock found in the upper portion of one sill. It is characterized by elongated albite crystals in a dark matrix of altered augite, chlorite, quartz, and fine-grained mesostasis. Plagioclase (albite) contrasts with that of other pegmatites (andesine). Chlorite is the same high-birefringent variety as that in other pegmatitic gabbros. Ores have a platy habit characteristic of this rock and are in amounts up to 11 per cent.

Average volumetric analyses for the rock types are given in Table 2 and are graphed in Figure 8. Pyroxene and plagioclase decline toward later members of the series, olivine decreases rapidly, iron-titanium oxides increase steadily throughout, and quartz appears in substantial quantities in the later members. In placing albite gabbro pegmatite on the end of the series the writer does not wish to imply that it was derived from gabbro pegmatite. It probably arose from the normal magma by a parallel process of differentiation and may be equivalent to gabbro pegmatite or subpegmatitic gabbro.

Leopard Rock

Sills of leopard rock may have three parts: border gabbro, coarse-grained leopard rock, and interior gabbro. The interior gabbro is probably a later intrusion and is not a necessary accompaniment of leopard rock.

BORDER PHASES: The border phases are similar to normal even-grained gabbro. Labradorite laths and equidimensional augite are ophitic or subophitic. Rare serpentine blebs are pseudomorphs after olivine. Iron ores form altered masses in interstices or show subophitic texture with plagioclase. Augite is partly replaced by hornblende, and plagioclase is saussuritized. The first feldspar clots 2–10 feet inward from the chilled margin are generally small phenocrysts 2–5 mm in diameter or clusters of two or three phenocrysts. Farther from the contact the size of phenocrysts and a number in a cluster increase until the clots have a diameter of 2–3 cm. As marginal gabbro passes into leopard rock the number of feldspathic clots increases rapidly, crowding the gabbroic groundmass into interstitial spaces and finally eliminating it as a recognizable phase. The phenocrysts throughout are heavily saussuritized plagioclase.

Volumetric data of a fine-grained specimen (±0.2 mm) 2 feet from the contact are:

Plagioclase phenocrysts	2
Plagioclase	47
Pyroxene	30
Hornblende	14
Chlorite	1
Iron ores	5
Serpentine (after olivine)	1

Table 11 shows the chemical analysis and norm of a marginal gabbro (R163) and a chilled but altered contact specimen (R161).

COARSE-GRAINED PHASE: The large altered plagioclase aggregates (6–15 cm in diameter) of the leopard rock are set in a feldspar-rich groundmass of 0.5- to 3-cm grain size, which consists of loosely packed feldspar clots and interstitial pyroxene, hornblende, quartz, biotite, chlorite, and apatite. Pumpellyite was observed in one section. Clots may be single, nearly equidimensional plagioclase crystals (±1 cm) or clusters of several crystals (1–3 cm in diameter) and are altered to saussurite, sericite, and locally prehnite. In most places marginal zones of feldspar crystals or clots are unaltered; whole crystals more rarely remain unaltered. Plagioclase shows strong progressive zoning with an abrupt outer rim. Small, little altered plagioclase laths (0.5 mm) are found in interstices and may be ophitically enclosed by pyroxene or hornblende. Augite crystals up to 2 cm across fill areas between feldspar clots. Hornblende forms large single crystals or masses of minute, randomly oriented acicular crystals. At least part of the hornblende is derived from pyroxene. Altered iron ores ranging from ½ to 3 mm in diameter enclose small interstitial plagioclase crystals ophitically. Brown biotite forms fringes around the ores. Quartz is interstitial and apparently crystallized last. Two sections show minor amounts of a light-brownish mesostasis composed of vaguely defined finely crystalline quartz charged with minute, threadlike crystals and dusty material. Locally the mesostasis surrounds and embays larger crystals of clear quartz and corrodes plagioclase. Possibly the material is devitrified glass. Interstitial, fine-grained mixtures of amphibole, quartz, epidote, and unidentified dusty material in other thin sections may be of similar origin. Apatite is generally embedded in mesostasis.

In a few places feldspar clots of the groundmass are more widely separated, and the spaces between them are occupied by light-colored, feldspathic gabbro.

INTERIOR GABBRO: Interior gabbros are similar to normal gabbros and are also differentiated. Near the base of the interior gabbro in the sill just west of Youngren Lake is typical porphyritic gabbro with olivine pseudomorphs. Directly above, the gabbro (Table 11, specimen B345) is nonporphyritic but still has olivine pseudomorphs. Even-grained gabbro without olivine follows. Specimen R195 (Table 11) is from the upper portion of the same interior gabbro, 3 miles farther south. Subpegmatitic gabbro and gabbro pegmatite were found below the upper contact with leopard rock in sills south of Lac Villeneuve and west of Murdock Lake.

Near its contact with leopard rock the gabbro is generally highly altered. Plagioclase is saussuritized, and orthopyroxene and olivine are replaced by serpentine. In two specimens with orthopyroxene pseudomorphs, serpentine is accompanied by a bladed mineral, probably anthophyllite.

COMPOSITE NATURE OF SILLS: The map pattern (Pl. 1) and cross section (Fig. 4) of the most westerly leopard-rock sill are convincing evidence that some leopard-rock sills are composite. Lenses of leopard rock isolated and cut by interior gabbro are supporting evidence. Kavanagh (1954, Ph.D. thesis, Princeton Univ., p. 74–78), from work in the Hyland Lake area 30–40 miles to the south, also concluded that sills of leopard rock are composite.

This interpretation may account for some of the peculiarities of leopard rock. The brownish mesostasis, interpreted as divitrified glass, may be due to partial melting of leopard rock by the later gabbro intrusion. Clear rims of sodic plagioclase surrounding crystals and aggregates of saussuritized plagioclase may have resulted from reaction between crystals and partial melt or high-temperature solutions from the interior gabbro. Perhaps alteration of pyroxene to hornblende, uncommon in normal gabbros, can also be attributed to the action of heat and emanations from the later intrusion.

Leopard rock-gabbro sills show the essential features of composite bodies elsewhere (*cf.* Harker, 1904, p. 197–214; Krokström, 1932, p. 308; Tomkeieff and Marshall, 1935, p. 269–276; Oftedahl, 1957, p. 4–5); two or more parallel intrusions—or extrusions (Kennedy, 1931, p. 167–169)—meet along unchilled contacts. Commonly members of a composite body are of contrasting rock types, but in composite lava flows described by Kennedy (1931, p. 167) the members are mugearite and highly porphyritic mugearite. Similarly, in Ahr Lake the members are gabbro and highly porphyritic gabbro. The unchilled contacts indicate that adjoining members were simultaneously liquid, or that one followed so shortly after consolidation of the other that no chilling resulted. The writer assumes that leopard rock was nearly consolidated at the time of the gabbro intrusion, since lenses of leopard rock are cut and isolated by gabbro. The site for gabbro intrusion may have been determined by a weakly consolidated zone at the center of the sills.

If the leopard rock was intruded before the gabbro, the alteration of gabbro adjacent to leopard rock requires explanation. Possibly late, aqueous fluids from the leopard rock, initially driven outward by gabbro intrusion, returned as the gabbro cooled to invade and alter its margins.

Metagabbro

GENERAL TYPE: The metagabbros are normally composed of albite, epidote (may include clinozoisite and zoisite), chlorite, actinolite, and sphene. Textures are metamorphic, but locally shapes of former plagioclase laths are apparent in dusty aggregates of epidote and albite. Some actinolite forms shreddy-edged, equidimensional crystals up to 4 mm in diameter or small acicular crystals; most is bent or broken. Chlorite is associated with actinolite or forms separate patches. Sphene forms small granular patches, some of which have relic cores of iron ore. Volumetric analysis of one specimen (B71) is: albite 22, epidote 21, actinolite 48, chlorite 5, sphene 4. The chemical analysis and norm of B71 are given in Table 11.

MELANOMETAGABBRO AND LEUCOMETAGABBRO: Melanometagabbro megascopically appears to be entirely amphibole but actually consists of large (0.5–3 cm) interlocking, ophitic amphibole crystals crowded with randomly oriented, dusty albite-zoisite laths and scattered irregular patches of colorless to pale-green chlorite.

Amphibole is colorless to light green, and some single crystals change sharply from one to the other. Some crystals contain curious circular structures marked by slight discontinuities between amphibole within and outside the structure. The two are in optical continuity but show a slight color difference. Where circular structures are in contact, the amphibole passes from one structure to the next without interruption, and areas between the structures contain shreddy amphibole or chlorite wedges. These textures are interpreted

as augite pseudomorphs poikilitically enclosing olivine. Olivine was apparently abundant. Hess (1933, p. 392) notes that pyroxene and included olivine may be replaced by a single amphibole

TABLE 3.—VOLUMETRIC ANALYSES OF MELANOMETAGABBRO AND LEUCOMETAGABBRO

	R337	R471	R470	R469	R468	R293
Albite	13	3	9	4	5	6
Epidote and albite	20	13	53	52	56	63
Amphibole	53	52	21	26	34	18
Iron ores	2	1	1	1	2	1
Chlorite	12	31	16	17	3	12

R337. Melanometagabbro, lower portion of Valley Ponds sill

R471–R468. Specimens from lower to upper zone of Protecto Lake sill: R471, melanometagabbro; R470 and R469, transitional rocks; R468, leucometagabbro

R293. Leucometagabbro, upper portion of Valley Ponds sill

crystal which varies slightly in composition between the portions replacing each of these minerals. This is confirmed by the slightly different colors between amphibole of the circular structures and that of the main crystal. Volumetric data are given in Table 3. These are only estimates in view of the coarse grain of the amphibole.

Leucometagabbro owes its unusually light color to the high proportions of altered plagioclase and the light-colored amphibole which it contains.

In thin section the rock is a confused intergrowth of amphibole, epidote, albite, chlorite, and sphene. Primary shapes of plagioclase are partly preserved by epidote-albite intergrowths rimmed with clear albite, but textures of former ferromagnesian minerals have been obliterated. Portions of some thin sections consist of closely packed pseudomorphs of plagioclase almost to the exclusion of ferromagnesian minerals. Amphibole is a colorless to pale-green tremolite-actinolite and commonly forms large ophitic crystals of widely separated parts. Chlorite is colorless and occurs in scattered patches. Volumetric data are given in Table 3.

The gradation from melanometagabbro to leucometagabbro in the Protecto Lake sill involves variations in relative amounts of mafic and felsic minerals and color of ferromagnesian minerals. Volumetric analyses of specimens across the narrow gradational zone are given in Table 3.

Specimen R471 of melanometagabbro and R469 of leucometagabbro are stratigraphically 200 feet apart. The transition is rapid; R470 is petrographically equivalent to the upper zone specimen R468. This is not so apparent in outcrop; specimen R470 is dark, whereas the leucometagabbro is light. The color change is not obvious in the volumetric analyses and is partly due to variation in iron content of amphibole and probably chlorite.

MINERALOGY

Normal Gabbros and Basalts of Western Zones

Mineralogical data are presented in Table 4 and summarized in Figure 9. Optic angles of pigeonite were not determined accurately; most appeared to be nearly zero and hence were plotted on the diagram assuming maximum Ca content (about 10 per cent). The data are principally from normal gabbros since most of the basalts are too fine-grained.

OLIVINE: Fresh olivine, found only in porphyritic gabbro, is hyalosiderite and richer in iron than coexisting orthopyroxene. Ramberg and De Vore (1951) found that magnesian olivine richer in iron than their coexisting orthopyroxenes. The cross-over at about 35 molecular per cent fayolite is approximately the composition of the present olivines. The reason for the anomalous behavior of olivine here is unknown. However, olivines in the Okonjeje complex, South-west Africa (Simpson, olivine is richer in magnesium and iron-rich 1954b, p. 132), are similar. They are richer in iron than coexisting orthopyroxenes and retain this relationship throughout the differentiated series.

ORTHOPYROXENE AND PIGEONITE: Fresh orthopyroxenes occur in several porphyritic gabbros and one even-grained gabbro. All except the most magnesian orthopyroxene (Fig. 9) have both (001) and (100) exsolution lamellae; hence they have inverted from pigeonite (Poldervaart and Hess, 1951, p. 481). The most magnesian orthopyroxene contains rare lamellae parallel to (100) and must have crystallized directly from the magma when the temperature was below the pigeonite-orthopyroxene inversion curve. According to Hess (1941, p. 581–582) this should take place before the bulk composition of crystals separating from a differentiating magma falls below

1609, colu

1609, column 2, place line 20 from bottom between lines 27 and 28 from bottom

a magnesium:iron ratio of 70:30. Pyroxenes crystallizing above the inversion curve may show evidence of inversion in the orientation of their augite exsolution lamellae (Poldervaart and Hess, 1951, p. 481). Figure 10 (A, B, D) shows orthopyroxenes with two sets of lamellae.

Pigeonite is scarce and extensively altered. The β index was determined with difficulty on small fragments removed from slides. Only one crystal shows augite exsolution lamellae.

Orthopyroxene (or pigeonite) is commonly irregularly intergrown with augite (Fig. 10A, B, C). Some single crystals of augite contain several patches of orthopyroxene or pigeonite with common orientation. The b and c axes of orthopyroxene or pigeonite in such intergrowths coincide with those of augite. Exsolution lamellae in the lime-poor pyroxene are in optical continuity with adjacent augite (Fig. 10B). Even augite lamellae in pseudomorphs after lime-poor pyroxene are in optical continuity with an adjacent augite.

Since lime-poor pyroxenes are susceptible to alteration, most are pseudomorphs. The identity of many original crystals can be determined by the augite exsolution lamellae. Twinned crystals, recognized by the chevron pattern of strings of augite lamellae (Fig. 10E), are pigeonite pseudomorphs. If lamellae of a pseudomorph are in optical continuity with adjacent augite, the original host was probably intergrown with augite (Fig. 10A). The physical orientation with respect to augite crystallographic axes identifies the lamellae as parallel

TABLE 4.—MINERALOGY OF NORMAL GABBROS

	B279	B184	B154	B174	R551	R126	B295	R90C	R90A	B300	B345	B337	R90J	R90K
Plagioclase*														
α'*	1.563–1.566	1.563–1.570	1.570	1.564	..	..	..	..	1.559	1.557–1.562	..	1.555–1.564	..	1.551–1.563
An	68–73	66–80	80	67	68–80	40–65	..	71	58	55–63	..	45–65	45–60	45–65
Augite														
2V(+)	48	50	n.d.	50–56	47	50	45	49	47	47	48	48	45	n.d.
β	1.687	1.689	1.687	1.683	1.689	1.688	1.690	1.687	1.689	1.684	1.685	1.686	1.689	n.d.
composition														
Ca	40	42	..	41–47	40	41	38	40	39	39	40	40	38	..
Fe	14	15	..	11–9	16	15	18	14	16	12	12	13	17	..
Mg	46	43	..	48–44	44	44	44	46	45	49	48	47	45	..
Orthopyroxene														
2V(−)	n.d.	58	68	..	n.d.	..	..	..	..	..	..	..	..	..
β	1.691	1.694	1.692	..	1.698	..	..	..	..	..	..	..	..	..
Fe	24	27	25	..	29	..	..	..	..	..	..	..	..	..
Pigeonite														
β	..	..	..	..	..	..	..	1.691	..	1.688	..	..	1.688	..
composition														
Ca†	..	..	..	..	..	..	..	9	..	9	..	..	9	..
Fe	..	..	..	..	..	..	..	33	..	31	..	..	31	..
Mg	..	..	..	..	..	..	..	58	..	60	..	..	60	..
Olivine														
2V(−)	84	n.d.	77	..	..	..	..	..	..	..	..	..	..	..
β	1.720	1.725	1.721	..	..	..	..	..	..	..	..	..	..	..
Fa	32	35	33	..	..	..	..	..	..	..	..	..	..	..

* Where no index shown, plagioclase determined by universal stage methods

† 2V not determined accurately, Ca_9 assumed

Porphyritic gabbro: B279, B184, B154, B174, R551, R126

Ophitic gabbro: B295, R90C, R90A, B300, B345, B337

Even-grained gabbro: R90J, R90K

TABLE 4.—(*Continued*)

	B357	B358	B287	R127	R195	R197	R67A	R67	R147	B286	R23	R56	R129	R128
Plagioclase														
α'	1.564	1.564	..	..	..	..	..	..	..	..	..	..	..	1.534
An	67	67	43–55	47–57	53	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	45–53	38–48	14
Augite														
2V(+)	51	n.d.	46	50	52	51	52	45–52	46	48–51	48–59	52	55	53
β	1.695	n.d.	1.690	1.706	1.695	1.692	1.696	1.704	1.690	1.694	1.691	1.693	1.689	1.704
composi-														
Ca	42	n.d.	39	39	43	42	42	35–41	39	40–42	40–49	43	46	42
Fe	20	n.d.	17	31	19	17	20	31–28	17	20–19	18–14	18	14	27
Mg	38	n.d.	44	30	38	41	38	34–31	44	40–39	42–37	39	40	31
Ortho-														
pyroxene														
β	1.707	..	..	..	..	..	..	..	..	..	..	..	..	..
Fe	37	..	..	..	..	..	..	..	..	..	..	..	..	..
Pigeonite														
β	..	..	1.692	..	..	..	..	..	..	..	..	..	..	..
compo-														
sition														
Ca	..	..	9	..	..	..	..	..	..	..	..	..	..	..
Fe	..	..	34	..	..	..	..	..	..	..	..	..	..	..
Mg	..	..	57	..	..	..	..	..	..	..	..	..	..	..

Even-grained gabbro: B357, B358, B287, R127, R195, R197, R67A
Subpegmatitic gabbro: R67, R147, B286, R23, R56, R129
Gabbro pegmatite: R128

	R554	R93	R45	B321	R97	R151	R83	R92	R94	B54A
Plagioclase										
α'	..	..	1.532	1.530	1.530	1.530	..	..	..	..
An	30–43	50	8	5	5	5	4	4	0	50–70
Augite										
2V(+)	55	42–47	46	46	n.d.	53	48	48	49	49
β	1.716	1.702	1.698	1.701	1.712	1.710	1.696	1.697	1.704	1.698
composition										
Ca	40	34–38	38	38	n.d.	41	39	39	39	39
Fe	39	29–27	24	27	n.d.	33	22	23	28	24
Mg	21	37–35	28	35	n.d.	26	39	38	33	37

Gabbro pegmatite: R554
Albite gabbro pegmatite (transitional): R93
Albite gabbro pegmatite: R45, B321, R97, R151, R83, R92, R94
Basalt: B54A

to (001) or (100). The original host can then be identified by applying the conclusions of Poldervaart and Hess (1951, p. 481).

Brown (1957, p. 542–543) has recently shown that orthopyroxene does not invariably retain the b and c axes of pigeonite upon inversion. Where these axes are not retained, augite lamellae exsolved on (100) of orthopyroxene will have no predictable relationship with the (001) lamellae of the parent pigeonite or with intergrown augite. Textures attributable to this type of inversion were found in two

porphyritic gabbro specimens (Fig. 10E). The chevron pattern of augite lamellae parallel to (001) of a former twinned pigeonite is crossed obliquely by a second set of augite lamellae. in augite are assumed to be pigeonite pseudomorphs. Their relationship with augite is markedly different from earlier members where augite and lime-poor pyroxene are intimately

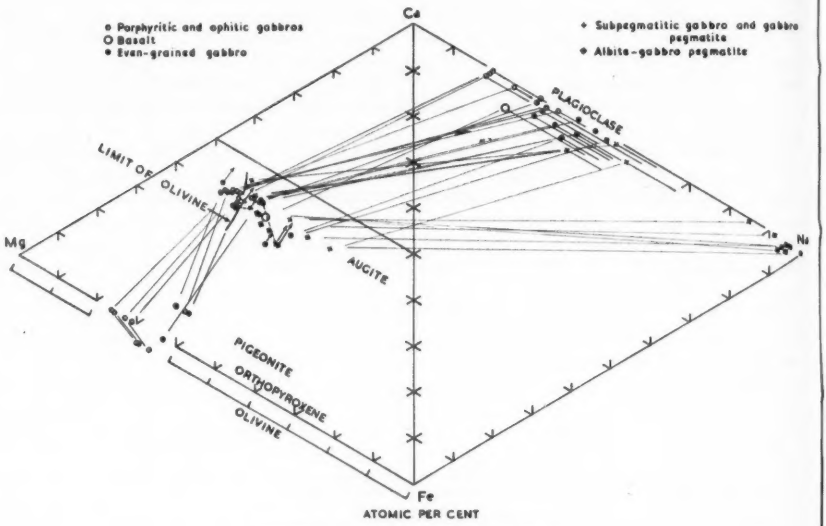


FIGURE 9.—MINERALOGICAL DIAGRAM

Variation in composition of principal minerals of normal gabbro series. Plagioclase compositions along the Na-Ca side are displaced short distances. Dashed lines join augites in contiguous even-grained and pegmatitic gabbros.

These are probably (100) lamellae of orthopyroxene but lie at an angle to the (100) twin composition plane of the parent pigeonite and cross it undiverted. Had the b and c axes been retained on inversion the second set of lamellae should parallel the twin. Brown (1957, p. 539) noted that nonretention of the b and c axes of lime-poor pyroxene on inversion becomes more prominent with height in the layered series of the Skaergaard intrusion. There is no evidence that this is true for successively later members of the normal gabbros in the Ahr Lake area.

Augite lamellae in orthopyroxene or pigeonite are confined to mafic gabbros. Few exsolution lamellae are found in rocks which do not also contain a few olivine pseudomorphs. Poldervaart and Hess (1951, p. 433) note that iron-rich pigeonites in hypabyssal rocks do not invert and exsolve exceedingly narrow augite lamellae on (001). Such lamellae may have been replaced, as well as the pigeonite host. In subpegmatitic gabbro small chlorite patches intergrown. They may have crystallized simultaneously in the early members, whereas in subpegmatitic gabbro pigeonite crystallized before augite.

AUGITE: Augite is present throughout and generally little altered. Figure 9 shows plots of composition variations; to the right of the line showing the limit of olivine association with augite olivine or olivine pseudomorphs are absent. The data are given in Table 4. A puzzling variation in optic angles was found in some specimens where 2V varies both from one crystal to the next and within individual crystals. Hotz (1953, p. 686) noted a similar variation of 2V in the Dillsburg, Pennsylvania, normal dolerite. The variation is not progressive as in zoning but changes abruptly from one portion of a crystal, not necessarily the center, to another (Fig. 11A). Portions of the crystal with the larger optic angle have a smoother surface (*i.e.*, less distinctly marked with cleavage and fractures) and lower bire-

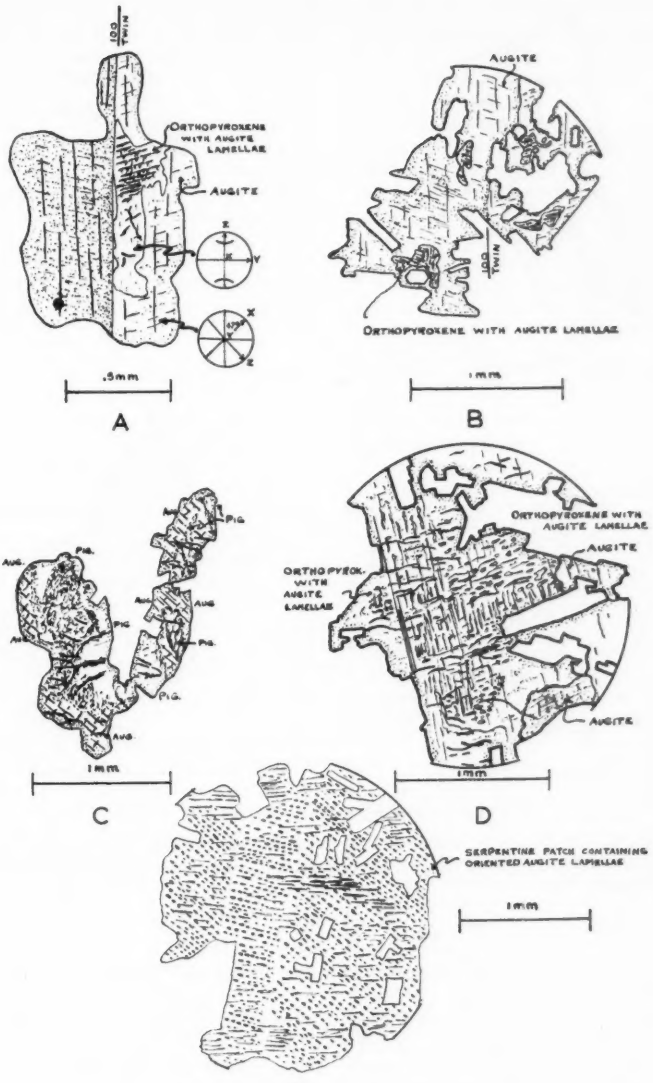


FIGURE 10.—PYROXENE INTERGROWTHS

A. Orthopyroxene-augite intergrowth. B. Disconnected orthopyroxene patches in augite. All are in optical continuity. C. Irregular pigeonite-augite intergrowths; b and c axes common to both. D. Orthopyroxene with augite lamellae. Associated augite patches are in optical continuity with augite lamellae. E. Serpentine after inverted pigeonite with chevron-patterned strings of augite lamellae showing former twinning. Elongated augite lamellae crossing twin plane are probably exsolved from orthopyroxene.

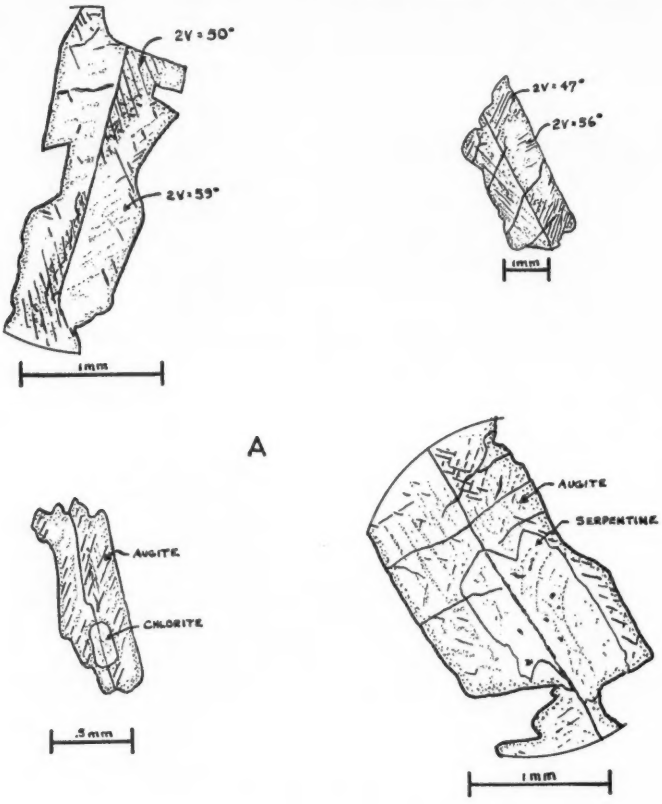


FIGURE 11.—AUGITE WITH VARIATIONS IN 2V AND CORES OF SERPENTINE AND CHLORITE
A. Augite crystals with areas of different optic angle. B. Augite with cores of serpentine and chlorite, possibly after pigeonite.

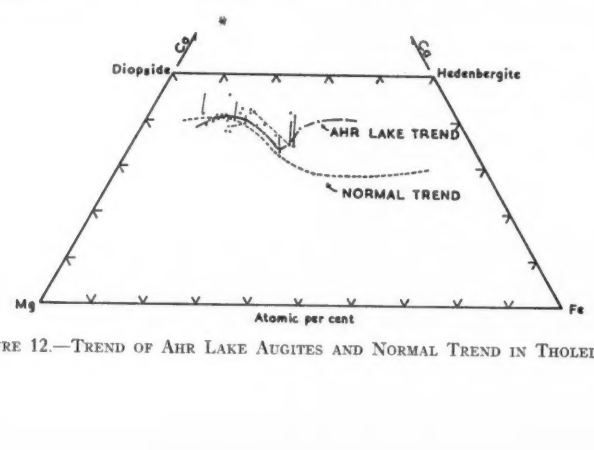


FIGURE 12.—TREND OF AHR LAKE AUGITES AND NORMAL TREND IN THOLEIITES

fringence. Were this portion of the crystal more calcic, as suggested by the larger optic angle, the birefringence should be higher, not lower. The low birefringence may be due to lamellar alteration along (001) and the large 2V to leaching of $FeSiO_3$ as suggested by Walker and Poldervaart (1949, p. 692) for pyroxene in weathered Karroo dolerite. Refractive index, determined in most cases on powders, showed no variation between particles of a sample within the limits of error (± 0.002). Optic-angle variations are therefore recorded on the diagram by a line and arrow along equal-index contours. If the variation is due to alteration, the base of the arrow records the actual optic angle.

The trend of Ahr Lake augites is compared in Figure 12 with the normal trend for tholeiites (Hess, 1941, p. 585). The Ahr Lake trend follows the normal trend initially but swings to higher calcium in late stages primarily because of augites from pegmatitic gabbros. Since Hess (1941) did not include augite from pegmatites the two curves are not comparable in this section. The higher volatile content of pegmatites would cause formation of a more ordered (*i.e.*, more calcic) augite by lowering the crystallization temperature of the magma (Poldervaart and Hess, 1951, p. 475). In addition a more sodic plagioclase would crystallize and leave additional calcium for augite. This does not appy in the first half of the course of crystallization where augites from pegmatites are along the normal trend from augites in contiguous even-grained gabbro (Fig. 12). Possibly at this stage of differentiation the pegmatitic phase is not sufficiently rich in volatiles to cause significant lowering of the crystallization temperature.

PLAGIOCLASE: Plagioclase in normal gabbros ranges from An_{80} to An_0. Many intermediate to calcic varieties are strongly zoned. Plagioclase in most of the sills shows a range from An_{80}–An_{60} in porphyritic gabbros to An_{55}–An_{35} in pegmatitic gabbros. In the sill with albite gabbro pegmatite, plagioclase of the pegmatite (An_0–An_8) differs markedly from plagioclase (An_{50}) in the transitional specimen (R93), immediately below the pegmatite. Walker's (1953, p. 46) conclusion that plagioclase of gabbro pegmatites is generally slightly more sodic than that of adjoining gabbro is true for most of the Ahr Lake gabbros. The composition gap in plagioclase from the sill with albite gabbro pegmatite suggests that this is not the result of normal differentiation.

Basalts contain labradorites,[3] except for three specimens which contain albite. These are not greatly altered and are similar to other basalts. Possibly they represent a limited development of spilite.

OPAQUE MINERALS: Sulfides are minor throughout the gabbros and slightly more abundant in the early differentiates. Pyrrhotite predominates in porphyritic and ophitic gabbros. Chalcopyrite and rarely pyrite are found as small, interstitial blebs.

Iron ores are most commonly ophitic masses of turbid, granular sphene enclosing triangularly or rectangularly arranged lamellae of ilmenite (Fig. 13A). The sphene was identified by X-ray powder photographs. Elongated, skeletal ilmenite crystals (Figs. 7B; 13B) are restricted to albite gabbro pegmatite. The sphene masses with ilmenite lamellae are interpreted as altered titaniferous magnetite with exsolved ilmenite. Replacement of magnetite by sphene while ilmenite remains unaltered seems surprising. Magnetite probably contained a fine meshwork of exsolved ulvöspinel (Vincent and Phillips, 1954, Pl. 1B), which weakened its resistance to replacement and provided the titanium necessary for sphene. Buddington *et al.* (1955, p. 501) observed sphene coronas on titaniferous magnetite due to deuteric alteration. The location of iron previously resident in magnetite is not clear. In mafic gabbros iron ores are minor, and iron released from magnetite could probably be taken up by chlorite (smectite-chlorite) replacing olivine and lime-poor pyroxene. In some pegmatitic gabbros the iron is accommodated in iron-rich chlorite and pumpellyite. Magnetite was found in only two specimens. In one it contained ilmenite lamellae and was extensively altered; in the other it was homogeneous and fairly fresh. Ilmenite with minor exsolved hematite is rare.

The oxide minerals form a guide to the state of oxidation in the magma (Buddington *et al.*, 1955) which in turn influences the trend of differentiation (Kennedy, 1955, p. 501; Wilkinson, 1957, p. 452). Alteration in the Ahr Lake gabbros makes interpretation of the oxide mineralogy speculative. If sphene represents earlier magnetite-ulvöspinel intergrowths the magmatic environment had low oxidizing potential, although the increase of oxide minerals from porphyritic to pegmatitic gabbro indicates that the potential was increasing

[3] Only plagioclase from the analysed basalt B54A is shown in Figure 9.

with differentiation. Ilmenite is more abundant relative to magnetite (sphene-ilmenite masses) in pegmatitic gabbros, especially albite gabbro pegmatite. This indicates more strongly oxidizing conditions in the later stages, if ulvöspinel was one of the original iron ores.

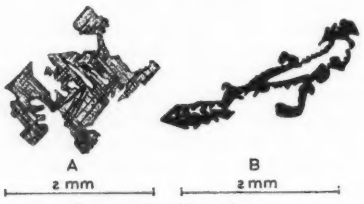


FIGURE 13.—IRON ORES

A. Rectilinear intergrowth of ilmenite in sphene. Probably after magnetite-ilmenite intergrowth. B. Skeletal crystal of ilmenite.

CHLORITE: A variety of chlorite with high birefringence and brownish to greenish pleochroism occurs in pegmatitic gabbros. Both birefringence and pleochroism become progressively more marked from mafic subpegmatitic gabbro to iron-rich gabbro pegmatites. Properties of an extreme variety from albite gabbro pegmatite (R94) are given in Table 5, as well as those of a less conspicuous variety from subpegmatitic gabbro (R56). These are compared with an analyzed iron-rich chlorite from a New Zealand spilite. The X-ray powder pattern of the Ahr Lake chlorite (R94) is identical with that of the New Zealand chlorite. Comparison in thin section[4] shows the New Zealand chlorite to be similar to the less extreme varieties of Ahr Lake chlorite (R56). Battey classifies his chlorite as brunsvigite. More conspicuous varieties of Ahr Lake chlorite cannot be classified on the optical correlation diagrams of Winchell and Winchell (1951, p. 385) or Hey (1954, p. 284) as the high birefringence places them outside the field. They are probably of the same series as Battey's chlorite but may be richer in iron.

PUMPELLYITE AND PREHNITE: Pumpellyite is pleochroic from pale greenish yellow to deep blue green. Its optical properties were not determined, but its identity was confirmed by X-ray powder photograph.

Prehnite has the following optical properties: colorless, moderate relief, lamellar habit, parallel extinction, length-fast, birefringence about 0.030, 2V large (±60°), positive. Owing to difficulty in separating it, X-ray confirmation was not obtained.

APATITE: Euhedral rods of apatite are common accessories in interstices of the pegmatic gabbros (Fig. 7B) and may be elongated (±1 mm). Many apatite crystals have pleochroic halos; some of the halos extend the length of the rod, others for only part of its length. Larsen and Keevil (1942) and Hutton (1947) recognized radioactivity in apatites. Recently Clark and Altschuler (1958) determined that the radioactivity may be due to tetravalent uranium proxying for calcium in fluorapatite.

Leopard Rock

PLAGIOCLASE: Plagioclase from the fine-grained border specimen (R163) is slightly zoned labradorite (An_{60}). In leopard rock it is strongly zoned and ranges from An_{70} to An_{38}. A narrow rim at the outer margin has compositions variously determined as An_{34}, An_{33}, and An_{24}. Plagioclase forming the large clots is An_{72} according to the norm of analysis R181 (Table 11). Normative plagioclase of leopard-rock groundmass is An_{63}, about the average of optical determinations on zoned crystals.

PYROXENE: Only augite was found in leopard rock and marginal phases. Alteration to amphibole is common and may obscure evidence of a second pyroxene. Augite of the border zone (R163) is $Ca_{40}Mg_{48}Fe_{12}$, and that of leopard rock ranges from $Ca_{41}Mg_{48}Fe_{11}$ to $Ca_{47}Mg_{41}Fe_{12}$. Both augites are in the region of the Ahr Lake trend where olivine is associated with augite. Augite in the border zone is accompanied by rare olivine pseudomorphs, but the leopard rock has only one questionable pseudomorph.

AMPHIBOLE: Coarse-grained amphibole is deeply colored from Z = blue green, Y = olive green, X = deep yellow green, to Z = grass green, Y = olive, X = pale yellow. It is probably calciferous hornblende. In addition, masses of minute, acicular, light-yellowish-green amphibole locally fill interstices of the rock. Amphibole replaces pyroxene and may be secondary.

OTHER MINERALS: Ore minerals are ophitic masses of sphene, 1–3 mm in diameter, containing ilmenite lamellae in triangular arrangement. Sphene probably is a pseudomorph after magnetite. The ores are approximately as abundant as in mafic phases of normal gabbros. Irregular halos of brown biotite locally fringe

[4] Thin sections of the New Zealand spilite from the Cambridge University collection.

ore minerals. Quartz is a minor constituent in some specimens and may contain apatite needles. Chlorite is also minor and generally after amphibole or pyroxene. Plagioclase

metagabbro and melanometagabbro are given in Table 6. Five are from specimens taken across the gradational zone from melanometagabbro (R472) to leucometagabbro.

TABLE 5.—PROPERTIES OF AHR LAKE CHLORITES AND IRON-RICH NEW ZEALAND CHLORITE*

	R94	R56	New Zealand chlorite		
Habit	Micaceous	Micaceous	..	SiO_2	27.11
Pleochroism:				TiO_2	0.35
x	Deep reddish brown	Pale yellowish brown	Pale yellow	Al_2O_3	17.42
				Fe_2O_3	2.91
y	Grass green	Pale green	Grass green	FeO	30.98
z	Grass green	Pale green	Slightly darker green	MnO	n.d.
				MgO	9.75
Refractive index:	..	..	..	CaO	0.21
γ	1.658 ± 0.002	1.639 ± 0.002	1.638	Na_2O	n.d.
α	1.641 ± 0.002	..	..	K_2O	n.d.
$\gamma - \alpha$	0.017	..	0.010 (approximately)	H_2O	11.07
2V	Essentially zero (−)	Essentially zero (−)	Small, (−)	H_2O	0.51
					100.31
Specific gravity	..	..	2.988		

* Battey (1956, p. 93)

alteration products form a fine-grained mixture in which individual minerals are difficult to identify. Sericite is generally present, and in this respect the alteration mixture differs from that of normal gabbro.

Metagabbro and Doublet Metabasalt

PLAGIOCLASE: Plagioclase rarely forms discrete crystals but is intimately intermixed with fine-grained epidote. Plagioclase fragments from a number of specimens of both metagabbro and metabasalt were determined as albite. In thin section it invariably has an index less than Canada balsam.

AMPHIBOLE: Amphiboles range from pale to dark green. In thin section most are colorless or pale green, but a few are darker green, blue green, or greenish brown. They are presumed to belong to the tremolite-ferrotremolite series.

Properties of amphiboles from leuco-

Specimen R469 is from the upper part of this zone. The iron content of the amphibole drops sharply as the upper part of the sill is approached. Amphibole in specimen R471 separated into heavy, dark-green and light, pale-green fractions. Possibly these correspond to amphiboles replacing olivine and pyroxene respectively (Hess, 1933, p. 392). Specimens R337 and R293 are from the lower and upper portions respectively of the Valley Ponds sill, and R235 is leucometagabbro of the same sill farther north where no underlying melanometagabbro is found. Iron content of amphibole drops from the lower to the upper part of this sill as well.

OTHER MINERALS: Chlorite is present in almost all specimens of metagabbro and metabasalt but is generally subordinate to amphibole. It is pale green, pleochroic, and shows anomalous blue or brown interference colors. Epidote forms gray, dusty clusters in which individual crystals are difficult to

identify. Less commonly clinozoisite and epidote are in larger, euhedral crystals. Sphene is minor but ubiquitous, and quartz is rare.

TABLE 6.—AMPHIBOLE PROPERTIES

	Rock type	Specific gravity	β	Color	Composition
R337	Melanometagabbro	3.13	1.635–1.646	α Colorless β Olive green γ Blue green	$Ferrotremolite_{20-33}$ Average $ferrotremolite_{30}$
R472*	Melanometagabbro	3.11	1.633–1.650	Colorless, Rarely pale Green	$Ferrotremolite_{19-38}$ Average $ferrotremolite_{27}$
R471* Light	Melanometagabbro	3.07	1.629	Colorless	$Ferrotremolite_{17}$
R471* Heavy	Melanometagabbro	Impure	1.648	α Pale green β Olive green γ Blue green	$Ferrotremolite_{35}$
R470*	Transitional	Impure	1.647	α Pale green β Olive green γ Blue green	$Ferrotremolite_{34}$
R469*	Transitional	Impure	1.635	Colorless, pale green	$Ferrotremolite_{20}$
R235	Leucometagabbro	3.11	1.642	Pale green	$Ferrotremolite_{28}$
R293	Leucometagabbro	Impure	1.632	Colorless	$Ferrotremolite_{18}$

* From same sill. Specimens progressive from lower portion R472 up

CHEMICAL DATA

General Statement

Chemical analyses and norms of igneous rocks of the Labrador trough are presented in Table 11. All but four are new analyses. Samples B279-R554 and B54A include analyses of normal gabbros and lava and show the range of differentiation. Spectrographic analyses, many from the same specimens, are presented with recalculated chemical data in Table 8. Nockolds and Allen's (1953; 1956) method of representation was adopted, and results can be compared directly.

Normal Gabbro Series

MAJOR ELEMENTS: Variations of the major elements are shown in Figures 14, 15, and 16.

Figure 14 shows the main features of differentiation in the Ahr Lake gabbros: a high degree of iron enrichment with minor enrichment in alkalies and exceptionally low potassium content throughout. The Skaergaard liquids (Wager and Deer, 1939, p. 218–219) are also shown for comparison. The similarity of the Ahr Lake trend with the Skaergaard line of descent is striking. The final stages are represented in the Skaergaard by various granophyres but are missing in the Ahr Lake series. Possibly the magmatic residue was dispersed through later members of the series. Since the magmatic residue of the Skaergaard is only about 5 per cent of the magma it would alter the iron-enrichment trend only slightly if dispersed. In the sodium-potassium-calcium diagram, the Ahr Lake series differs from the Skaergaard liquids; potassium remains low throughout, whereas in the Skaergaard it increases rapidly with differentiation.

Figure 15 illustrates the variation in normative minerals with differentiation. The Skaergaard trend is represented by rocks and confined to the layered series. The Ahr Lake trend fits that of the Skaergaard remarkably closely and falls only slightly short of its iron-rich extremity.

In Figure 16, the measure of differentiation used is the mafic index (Wager and Deer, 1939, p. 168; Simpson, 1954a, p. 238) which assumes that the stage of differentiation reached is indicated by the extent of iron enrichment; this is probably justified since the Ahr Lake series shows extreme iron enrichment with

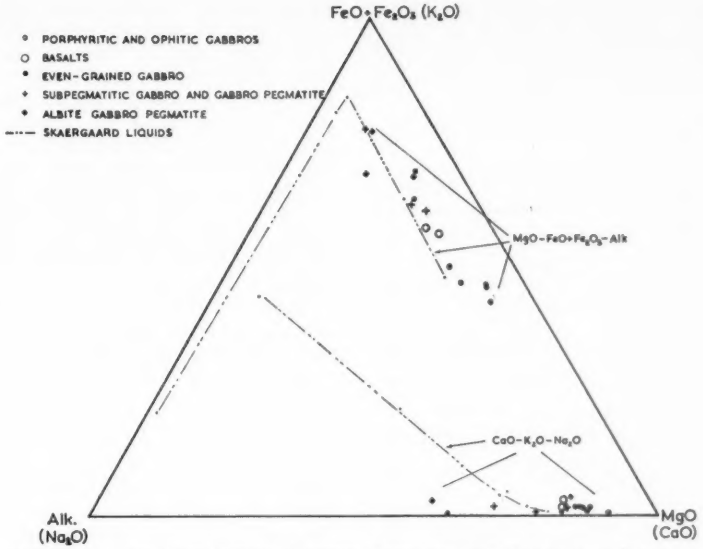


FIGURE 14.—AHR LAKE NORMAL GABBROS COMPARED WITH SKAERGAARD LIQUIDS

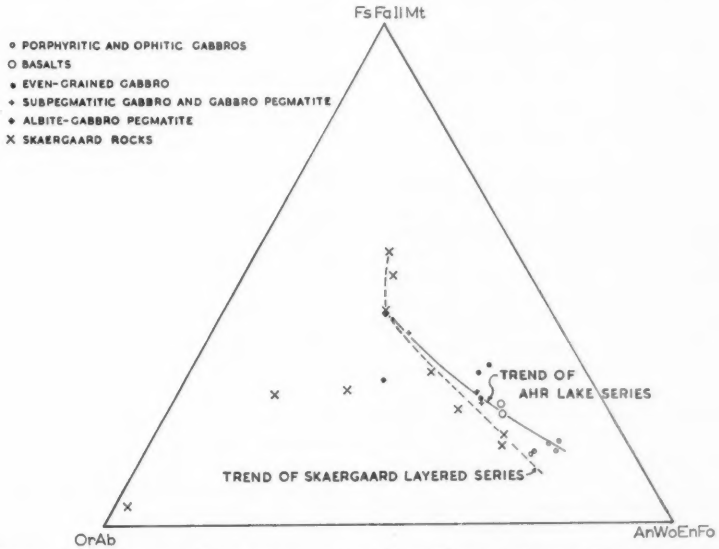


FIGURE 15.—NORMATIVE COMPONENTS OF AHR LAKE NORMAL GABBROS COMPARED WITH SKAERGAARD ROCKS

little change in alkalies. The main features of the diagram are: (1) iron oxides increase markedly; (2) silica and titanium oxide increase moderately and alkalies slightly; (3) magnesium, calcium, and aluminum oxides decrease, calcium more rapidly in the later stages, aluminum in the early stages, and magnesium steadily throughout.

The chemical variations reflect the changing mineralogy throughout the series (Fig.8). Alumina declines as plagioclase decreases and its albite content rises. Silica increases steadily

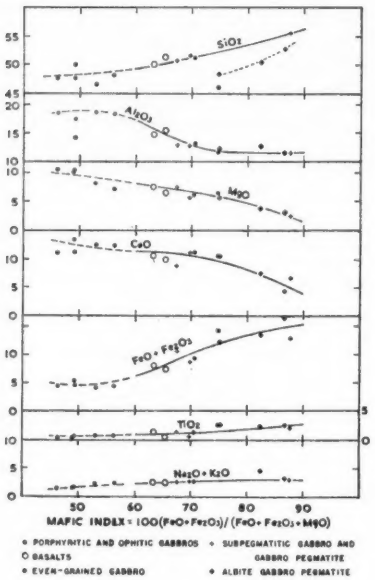


FIGURE 16.—VARIATION OF OXIDES WITH MAFIC INDEX IN NORMAL GABBROS

owing to an early decline in olivine followed by a rise in free quartz. The lime curve results from variations in quantity and composition of both augite and plagioclase. Probably the initial slow decline in lime reflects decreasing anorthite nearly balanced by slight increase in augite. In later stages, both decrease and the lime curve turns more steeply downward. The marked increase of iron is reflected by increasing iron content of silicates and steady rise of iron-titanium oxides. Alkalies increase slowly throughout the series; hence there is a near balance between decreasing plagioclase and rising albite content.

MINOR ELEMENTS: Variations in major and minor elements of the normal gabbro series (Table 8) are shown in Figure 17. The diagram is the modified Larsen diagram used by Nockolds and Allen (1953; 1954; 1956) in which the abscissa is the function (⅓ Si + K) − (Ca + Mg) and the ordinate the metal in parts per million or per cent. Table 8 contains two sets of spectrographic analyses. Partial analyses were done at the Lamont Geochemical Laboratory using Turekian's quantitative method (Turekian and Kulp, 1956). More nearly complete analyses were later made at Cambridge, England, on a larger suite of rocks, using the semiquantitative method of Nockolds and Allen (1953, p. 105-106). The same powders were used; hence differences in results should reflect differences in techniques. The Cambridge results are generally 30–50 per cent lower than those from Lamont. Only the Cambridge results are graphed in Figure 17.

The minor elements show an erratic distribution about the curves, which may reflect slight differences in differentiation from sill to sill. Following is a summary of the behavior of the minor elements:

Gallium remains nearly constant or increases slightly. According to Wager and Mitchell (1951, p. 182) and Rankama and Sahama (1950, p. 720–721), it substitutes for aluminum in silicates and ferric iron in magnetite. In Nockolds and Allen's (1956) diagrams gallium remains constant or declines slightly.

Chromium is abundant in early members of the series where it may replace ferric iron in augite and magnetite (Wager and Mitchell, 1951, p. 183–184) but declines rapidly in the middle stages of the series. This is the characteristic behavior in tholeiitic series studied by Nockolds and Allen (1956, p. 47, 51–55).

Vanadium distribution is erratic. Generally it is high in the middle stages and disappears in the late stages. Nockolds and Allen (1956) show vanadium concentrated in early or middle phases of tholeiitic series and declining to zero in the later stages; thus its behavior here is similar. Wager and Mitchell (1951, p. 162) show that it is probably present in magnetite and ilmenite.

Lithium distribution is erratic, but in general it declines with increasing values of the modified Larsen function. Its ratio with magnesium increases. In tholeiitic series elsewhere (Nockolds and Allen, 1956, p. 47, 51–55) lithium remains nearly constant or declines slightly with differentiation, but the lithium: magnesium ratio increases.

Nickel is concentrated in early members of the series and decreases rapidly and disappears in the middle stages. This is its characteristic behavior in igneous series (Nockolds and Allen, 1953; 1954; 1956; Wager and Mitchell, 1951), where it substitutes for magnesium or ferrous iron in early formed olivines and to a

lesser extent in pyroxene and iron-titanium oxides (Wager and Mitchell, 1951, p. 187). In the present series it is probably largely in serpentine pseudomorphic after olivine or orthopyroxene and in augite.

According to Rankama and Sahama (1950, p. 565–566) and Mason (1952, p. 115–116), zirconium fits with difficulty into the common rock-forming minerals; hence it is present in igneous rocks principally as zircon. However,

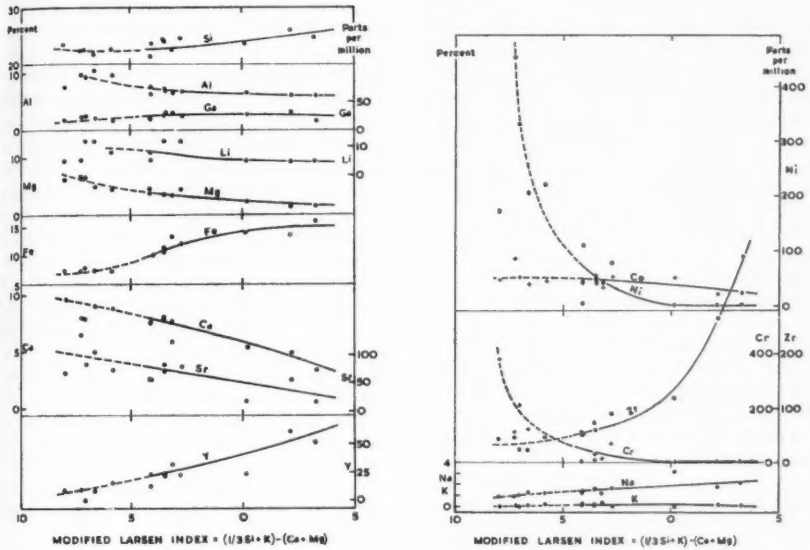


FIGURE 17.—VARIATION OF MAJOR AND MINOR ELEMENTS OF NORMAL GABBRO SERIES

Cobalt is constant throughout most of the series but declines slightly in the late stages. The cobalt:nickel ratio increases. This behavior is similar to some of the tholeiitic series studied by Nockolds and Allen (1956, p. 54), particularly the Dillsburg sill. According to Wager and Mitchell's (1951, p. 162) data, cobalt is distributed among pyroxene, ilmenite, magnetite, and olivine in proportion to their abundance in the rock. Like nickel, it substitutes for magnesium and ferrous iron.

Scandium is erratic. In the Skaergaard scandium is in pyroxene and apatite, where it substitutes for calcium (Wager and Mitchell, 1951, p. 192). It rises to a maximum in the hortonolite ferrogabbro and declines in later differentiates. Scandium is lowest in the early and late members of the Ahr Lake series, and to this extent its distribution is similar to that of the Skaergaard. This probably reflects similar pyroxene distribution.

Zirconium increases exponentially from earliest members of the series to the latest. Wager and Mitchell (1951, p. 192) found zirconium in pyroxene and apatite where it substitutes for calcium. Zircon was not observed in the Ahr Lake gabbros but may be present in the pegmatitic rocks. Otherwise zirconium is presumably in pyroxenes and plagioclase and in apatite in the late stages of the series. Its increasing concentration throughout the series reflects the difficulty with which zirconium fits into structures of common rock-forming minerals. This appears to be its normal behavior in tholeiitic series as shown by Nockolds and Allen (1956).

Yttrium increases steadily throughout. This is its normal behavior in tholeiitic series according to Nockolds and Allen (1956). It occurs in apatite, sphene, and pyroxene (Wager and Mitchell, 1951, p. 163; Rankama and Sahama, 1950, p. 524), where it replaces calcium. In the Ahr Lake rocks it probably occurs in augite and in later stages also in apatite.

Strontium shows a steady decline with increasing values of the modified Larsen func-

tion. This differs from its normal behavior in tholeiitic series (Nockolds and Allen, 1956, p. 61–65), where the strontium content remains nearly constant or increases slightly with rising values of the differentiation function. However, in common with its behavior in these series, the strontium:calcium ratio increases with differentiation throughout. Strontium follows calcium, but Turekian and Kulp (1956, p. 265) demonstrated that in basaltic rocks its concentration is independent of the amount of calcium, so that its variation is principally due to magmatic differentiation or initial regional differences. During differentiation, strontium is concentrated relative to calcium, and this is shown in the Ahr Lake series by the rising strontium:calcium ratio. The limited spread of strontium values because of differentiation is generally outweighed by more marked regional differences between provinces (Turekian and Kulp, 1956, p. 267–269). The average strontium content of the Ahr Lake gabbro series, 83 ppm (using Lamont analyses), is lower than that of the 30 provinces listed by Turekian and Kulp (1956, p. 268).

Barium remains at a low and fairly constant amount throughout the series, with the exception of two erratically high values. Since barium follows potassium its behavior reflects the low potassium content of the gabbros.

Rubidium also follows potassium and is extremely low in the series.

Copper and tin were determined at Lamont on partial series. The limited information obtained shows no systematic variation in either element. Copper is probably present in sulfides. Tin may be present as a replacement of calcium or ferrous iron in silicates (Rankama and Sahama, 1950, p. 732).

Metagabbros

MAJOR ELEMENTS: Six analyses of metagabbro (Table 11) include one each of melanometagabbro (R337), leucometagabbro (R293), and blotchy metagabbro (202–50). Remaining analyses are of ordinary metagabbros,[5] and their average is compared in Table 7 with average normal gabbro. The most significant difference between the averages is a substantially lower iron-magnesium ratio in metagabbro, despite higher silica and alkalies. Probably the metagabbros belong to the same province as the normal gabbros, but their compositions may have changed during metamorphism. The composite sample of Doublet metabasalt (Table 11) fits the normal gabbro variation diagrams closely and is similar in composition to basalt of the western zones; hence the metagabbro abnormalities may be local and might disappear in larger samples.

[5] Analyses 46–50, 121–50, and 202–50 are unpublished analyses from the Griffis Lake area generously provided by W. F. Fahrig. Blotchy metagabbro (202–50) is equivalent to metagabbro with feldspar clots associated with blotchy metabasalts in the Ahr Lake area.

TABLE 7.—METAGABBROS AND OTHER ANALYSES
(Recalculated water-free, to 100 per cent)

	1	2	3	4	5	6
SiO_2	52.12	49.74	47.90	47.50	48.05	45.89
TiO_2	0.67	1.56	0.39	0.70	0.19	1.81
Al_2O_3	12.83	14.47	23.52	14.56	24.60	15.36
Fe_2O_3	2.45	1.90	1.45	1.16	2.30	1.22
FeO	9.75	12.45	4.32	9.14	1.80	8.12
MnO	0.14	0.25	0.07	0.18	0.05	0.08
MgO	8.76	6.86	6.75	15.21	8.71	13.22
CaO	9.67	10.20	12.78	9.19	11.17	10.71
Na_2O	2.86	2.20	2.72	2.04	2.80	2.30
K_2O	0.42	0.22	0.10	0.06	0.28	0.67
P_2O_5	0.14	0.15	tr.	0.26	0.05	0.62
Rest	0.19	..	..	..	..	..

1. Average metagabbro (3 analyses)
2. Average normal gabbro (14 analyses)
3. Leucometagabbro R293, Ahr Lake area
4. Melanometagabbro R337, Ahr Lake area
5. Saussurite-smaragdite gabbro, moraine of Allalin glacier near Zermat, Switzerland (Burri and Niggli, 1945, p. 597)
6. Olivine basalt lava north slope Hualalai, Hawaii (Washington, 1923, p. 102)

Leucometagabbro and melanometagabbro have unusual compositions. Melanometagabbro resembles olivine-rich basalts (Table 7), but the closest match for leucometagabbro is a saussurite-smaragdite gabbro from the Swiss Alps (Table 7). Parts of this rock resemble leucometagabbro, according to descriptions by Bonney (1892). Norms of melano- and leucometagabbro (Table 11) confirm that the former was rich in olivine and the latter was predominantly plagioclase.

MINOR ELEMENTS: Spectrographic analyses of three specimens of metagabbros and eight of melano- and leucometagabbros are given in Table 8. The metagabbros show little similarity to the normal gabbro series except for unusually low strontium values. The analy-

ses fit poorly the trace-element curves of normal gabbros.

The minor elements show anomalous behavior in sills with melanometagabbro and leucometagabbro. In a series of specimens from melanometagabbro to leucometagabbro in the Protecto Lake sill (R472–R468), chromium increases steadily, whereas normally it is concentrated in the ferromagnesian-rich fractions and diminishes toward the felsic fraction. Nickel behaves normally, with high values in melanometagabbro decreasing toward leucometagabbro, but vanadium ranges anomalously from high values in leucometagabbro to low values in melanometagabbro. The behavior of other trace elements is compatible with the changes in mineralogy. Strontium and gallium are both higher in leucometagabbro, whereas lithium is lower. Variation of trace elements from melanometagabbro (R337) to leucometagabbro (R293) in the Valley Ponds sill confirms the pattern found in the Protecto Lake sill, except that chromium is about the same in mafic and salic portions. Barium, zirconium, and scandium all show marked increases in the uppermost specimen of the Protecto Lake sill but are approximately the same in upper and lower parts of the Valley Ponds sill.

Leopard Rock

MAJOR ELEMENTS: Chemical analyses of leopard rock in Table 11 include the chilled contact (R161), the border zone of gabbro 2 feet from the contact (R163), the groundmass of fully developed leopard rock (R177), and one of the larger feldspathic clots (R181). The chilled contact has been metasomatized. Potassium is abnormally high compared with normal gabbro and other leopard-rock analyses, and lime is low. The norm of the feldspathic clot (R181) contains about 6 per cent nepheline, but if muscovite is substituted for orthoclase and zoisite for anorthite, the norm has minor quartz. Leopard rock resembles anorthositic gabbros (Buddington, 1939, p. 36), except for its lower silica.

MINOR ELEMENTS: Spectrographic analyses were made on specimens taken at intervals from the contact through the border zone (R163–172) into fully developed leopard rock (R175). All but the last are gabbro specimens from which feldspathic clots have been excluded. Significant features of the analyses are: (1) Minor-element content does not change systematically from the contact inward to leopard rock. Compositions of the four border-zone specimens are similar and differ from the leopard rock. This supports petrographic observations that gabbro of the border zone is uniform throughout, except for addition of feldspathic clots. (2) Strontium values in the border-zone gabbro are similar to those in normal gabbro and are exceptionally low compared with other basaltic provinces (Turekian and Kulp, 956, p. 268). Since strontium content characterizes a basaltic province (Turekian and Kulp, 1956, p. 267–269), a genetic relationship is suggested between border-zone gabbros of leopard-rock sills and the normal gabbro series.

METAMORPHISM

Regional Metamorphism

Metamorphism increases from west to east across the Ahr Lake area in three distinct zones separated by faults. The western zone shows such slight metamorphic effects that they cannot be distinguished with certainty from the effects of deuteric, diagenetic, or hydrothermal alteration.

QUARTZ-ALBITE-EPIDOTE-BIOTITE SUBFACIES: Biotite schists of the metamorphic complex east of the Labrador trough are poorly exposed in the northeast corner of the Ahr Lake area. Only metasedimentary rocks were found. A typical mineral assemblage is quartz-albite-biotite-chlorite-sphene±epidote. Tourmaline and magnetite are accessories. Biotite and chlorite form discrete crystals apparently in equilibrium with their environment. Bedding is well preserved locally. The rocks have pronounced foliation marked by parallel alignment of biotite and slight flattening of quartz and albite. The sodic plagioclase and the mineral association identify the metamorphism as quartz-albite-epidote-biotite subfacies of the greenschist facies (Fyfe *et al.*, 1958, p. 218, 223). Elsewhere similar biotite schists on the east side of the Labrador trough increase in metamorphism eastward (Fahrig, 1951; 1956a).

QUARTZ - ALBITE - MUSCOVITE - CHLORITE SUBFACIES: Rocks of this zone include basalts and basaltic pyroclastic rocks, argillitic and probably tuffaceous sediments, gabbros, and ultramafic sills.

The subfacies begins abruptly at the lineament formed by Murdock and Irene lakes and extends to Keato Lake. The mineral

TABLE 8.—GEOCHEMICAL DATA

Expressed as parts per million

	Normal Gabbro Series													
	R90C	B279	B295	R189	B345	R195	B54A	B286	R127	R93	R129	R151	R554	R94
Si ($\times 10^{-3}$)	234	222	223	217	226	214	234	242	240	225	247	234	260	246
Al ($\times 10^{-3}$)	76	98	92	104	98	63	78	72	70	66	68	67	61	60
Fe''' ($\times 10^{-3}$)	8	10	10	9	11	9	10	15	16	17	20	30	10	14
Fe'' ($\times 10^{-3}$)	66	63	70	63	62	142	91	92	98	118	103	112	129	148
Ca ($\times 10^{-3}$)	97	79	80	91	90	77	75	81	79	76	64	53	49	32
Mg ($\times 10^{-3}$)	60	67	65	49	44	39	46	36	37	34	46	24	15	19
Na ($\times 10^{-3}$)	10	10	11	14	14	13	15	17	18	14	18	33	20	22
K ($\times 10^{-3}$)	0.5	2	1	2	3	3	2	2	1	4	1	0.1	2	2
Ti ($\times 10^{-3}$)	5	4	5	5	5	18	8	5	8	18	9	14	13	16
P	228	1000	228	131	524	262	383	480	0	1310	0	436	1140	2700
Mn	1800	1410	1490	1490	1800	2120	1800	2120	1880	2280	2040	2120	2900	1960
Ga	18	13	15	10	7	10	10	17	21	20	15	17	24	7
Cr	750	220	425	235	21	5	180	49	7	20	130	..	..	..
Cr*	1000	145	550	390	410	29	320	76	..	..	..	..	..	17
V	245	195	250	220	225	1000	450	410	450	350	450	625	..	..
Li	5	5	12	12	8	10	5	12	12	n.d.	12	5	5	5
Ni	170	500	400	205	220	5	110	47	40	40	77	..	3	4
Ni*	177	335	480	184	330	28	112	81	..	..	..	..	..	15
Sn*	40	24	25	27	15	11	15	8	..	..	..	..	..	14
Cu*	130	104	130	71	150	44	145	128	..	..	..	..	..	120
Co	48	85	50	32	48	40	45	48	58	32	52	52	22	26
Sc	46	..	30	12	12	47	55	46	54	26	54	35	24	..
Zr	45	48	25	23	48	52	57	75	62	170	92	120	267	380
Y	10	10	..	8	16	12	24	24	22	32	23	23	62	52
Sr	70	150	80	105	72	57	57	83	70	125	82	18	57	24
Sr*	53	108	96	113	99	64	56	101	..	..	..	..	..	61
Ba	20	36	50	50	22	23	25	107	27	160	35	105	22	49
Rb	..	10	..	10	10	10	..	..	..	n.d.	..	..	..	..
Modified Larsen function	−7.9	−7.2	−7.0	−6.6	−5.8	−4.1	−4.1	−3.5	−3.5	−3.16	−2.7	+2.7	+2.2	+3.3

	Leopard Rock					Metagabbros								
	R90A	R90B	R90D	R90E	R90F	R90G	R90H	R90I	R90J	R90K	R90L	R90M	R90N	R90O
Si ($\times 10^{-3}$)	234	234	234	234	234	234	234	234	234	234	234	234	234	234
Al ($\times 10^{-3}$)	76	76	76	76	76	76	76	76	76	76	76	76	76	76
Fe''' ($\times 10^{-3}$)	8	8	8	8	8	8	8	8	8	8	8	8	8	8
Fe'' ($\times 10^{-3}$)	66	66	66	66	66	66	66	66	66	66	66	66	66	66
Ca ($\times 10^{-3}$)	97	97	97	97	97	97	97	97	97	97	97	97	97	97
Mg ($\times 10^{-3}$)	60	60	60	60	60	60	60	60	60	60	60	60	60	60
Na ($\times 10^{-3}$)	10	10	10	10	10	10	10	10	10	10	10	10	10	10
K ($\times 10^{-3}$)	0.5	0.5	0.5	0.5	0.5	0.5	0.5	0.5	0.5	0.5	0.5	0.5	0.5	0.5
Ti ($\times 10^{-3}$)	5	5	5	5	5	5	5	5	5	5	5	5	5	5
P	228	228	228	228	228	228	228	228	228	228	228	228	228	228
Mn	1800	1800	1800	1800	1800	1800	1800	1800	1800	1800	1800	1800	1800	1800
Ga	18	18	18	18	18	18	18	18	18	18	18	18	18	18
Cr	750	750	750	750	750	750	750	750	750	750	750	750	750	750
Cr*	1000	1000	1000	1000	1000	1000	1000	1000	1000	1000	1000	1000	1000	1000
V	245	245	245	245	245	245	245	245	245	245	245	245	245	245
Li	5	5	5	5	5	5	5	5	5	5	5	5	5	5
Ni	170	170	170	170	170	170	170	170	170	170	170	170	170	170
Ni*	177	177	177	177	177	177	177	177	177	177	177	177	177	177
Sn*	40	40	40	40	40	40	40	40	40	40	40	40	40	40
Cu*	130	130	130	130	130	130	130	130	130	130	130	130	130	130
Co	48	48	48	48	48	48	48	48	48	48	48	48	48	48
Sc	46	46	46	46	46	46	46	46	46	46	46	46	46	46
Zr	45	45	45	45	45	45	45	45	45	45	45	45	45	45
Y	10	10	10	10	10	10	10	10	10	10	10	10	10	10
Sr	70	70	70	70	70	70	70	70	70	70	70	70	70	70
Sr*	53	53	53	53	53	53	53	53	53	53	53	53	53	53
Ba	20	20	20	20	20	20	20	20	20	20	20	20	20	20
Rb	..	..	..	..	..	..	..	..	..	..	..	..	..	..
Modified Larsen function	−7.9	−7.9	−7.9	−7.9	−7.9	−7.9	−7.9	−7.9	−7.9	−7.9	−7.9	−7.9	−7.9	−7.9

	Leopard Rock					Metagabbros								
	R163	R165	R169	R172	R175	B71	R235	R293	R337	R472	R471	R470	R469	R468
Si ($\times 10^{-3}$)	234	..	..	..	..	253	..	223	222	..	..	..	..	..
Al ($\times 10^{-3}$)	89	..	..	..	..	74	..	124	73	..	..	..	..	..
Fe‴ ($\times 10^{-3}$)	6	..	..	..	..	7	..	10	8	..	..	..	..	..
Fe″ ($\times 10^{-3}$)	101	..	..	..	..	63	..	34	71	..	..	..	..	..
Ca ($\times 10^{-3}$)	72	..	..	..	..	57	..	91	65	..	..	..	..	..
Mg ($\times 10^{-3}$)	24	..	..	..	..	55	..	41	92	..	..	..	..	..
Na ($\times 10^{-3}$)	17	..	..	..	..	30	..	22	15	..	..	..	..	..
K ($\times 10^{-3}$)	4	..	..	..	..	1	..	1	0.5	..	..	..	..	..
Ti ($\times 10^{-3}$)	10	..	..	..	..	6	..	2	4	..	..	..	..	..
P	350	..	..	..	..	393	..	tr.	114	..	..	..	..	..
Mn	1570	..	..	..	..	1720	..	548	1410	..	..	..	..	..
Ga	15	12	15	15	24	5	15	17	5	5	5	12	12	17
Cr	120	105	117	200	115	245	190	457	510	160	180	250	505	550
Cr*	197	145	200	227	155	385	..	..	820	..	..	..	..	..
V	500	500	450	450	375	220	185	120	90	67	50	120	90	220
Li	12	25	12	8	12	12	12	8	15	12	12	5	5	5
Ni	170	220	110	210	122	190	30	190	550	435	435	210	240	70
Ni*	100	175	125	183	92	163	..	..	570	..	..	..	..	..
Sn*	13	14	15	12	19	18	..	..	33	..	..	..	..	..
Sn	..	..	..	..	..	..	15	15	..	10	10	10	10	22
Cu*	142	165	105	124	112	16	..	..	45	..	..	..	..	..
Co	52	68	45	50	49	45	29	32	53	49	62	40	52	30
Sc	55	28	33	40	28	46	46	14	12	..	..	12	10	47
Zr	100	92	100	87	51	52	42	26	24	26	27	28	26	51
Y	22	25	21	21	16	15	14	..	..	..	..	..	..	12
Sr	72	92	93	70	115	27	380	425	58	155	75	540	310	500
Sr*	80	108	118	89	155	44	..	..	103	..	..	..	..	..
Ba	50	47	36	28	47	11	338	12	12	28	35	23	17	125
Rb	25	32	14	..	18	..	32	..	..	..	..	..	..	..

* Analyzed at Lamont Geological Observatory; remainder analyzed at Cambridge, England

assemblage of each rock type has little systematic variation over this distance. The typical mineral assemblage of metagabbros and metabasalts is albite-epidote-actinolite-chlorite-sphene. Chlorite is subordinate to actinolite, minor biotite is present, and quartz is rare in most of the rocks. The mineral assemblage of the mafic rocks assigns them to the greenschist facies; further subdivision depends on mineral assemblages of associated argillites (Fyfe *et al.*, 1958, p. 218). Interbedded shales and siltstones show assemblages as follows: (1) albite-epidote-actinolite-chlorite; (2) quartz-albite-white mica-chlorite; (3) quartz-albite-white mica-chlorite-epidote; (4) quartz-albite-white mica-actinolite; (5) quartz-albite-white mica-biotite-chlorite. Sphene is ubiquitous, and tourmaline and opaque minerals are common accessories. Assemblages (2) and (3) occur most commonly and are responsible for classifying rocks of this zone in the quartz-albite-muscovite-chlorite subfacies. The association of white mica and chlorite in place of biotite is the main criterion for distinguishing this subfacies from the quartz-albite-epidote-biotite subfacies (Fyfe *et al.*, 1958, p. 223). Assemblage (5) is present in two specimens and contains both white mica-chlorite and biotite, but in each case biotite is partly replaced by chlorite. The significance of biotite is uncertain. Since most of the sedimentary rocks are profusely intruded by sills, biotite may be a relic of contact metamorphism. Assemblages (1) and (4) resemble those of metagabbro and metabasalt and probably represent tuffaceous sediments.

Schistosity is rare except in the Murdock formation where it may have resulted from movement on the subjacent fault. Elsewhere white mica, chlorite, or biotite may parallel bedding planes in metasedimentary rocks, but in metagabbros and metabasalts the minerals generally show random orientation.

SUBGREENSCHIST FACIES METAMORPHISM: Rocks west of the Murdock-Irene lakes valley show primary igneous and sedimentary textures and are composed at least partly of primary minerals. They cannot be considered a metamorphic facies, which requires that "... minerals of any particular assemblage show no textural or other evidence of mutual replacement" (Fyfe *et al.*, 1958, p. 18), but they may be expected to show some effects of regional metamorphism.

The sedimentary rocks are nearly unaltered. In wackes the grain boundaries are sharply defined and show no evidence of recrystallization. Potassium feldspar and plagioclase are clear or turbid. No new minerals were observed. Dolomites are generally fine-grained (less than 0.2 mm) and contain no new minerals. Iron formation should be a sensitive indicator of metamorphism because of the presence of chert. James (1955, p. 1474–1480) demonstrated the sensitivity of grain size to metamorphism in the iron formations of Michigan. In the chlorite zone he finds a grain size 0.03–0.05 mm. Iron formation near Hurst Lake contains chertlike layers with average grain size about 0.005 mm. Other layers are coarser-grained (up to 0.1 mm), but the rock as a whole must have undergone little metamorphism.

Gabbros show appreciable alteration, which increases through succeeding stages of differentiation from porphyritic gabbro to gabbro pegmatite. This is common to many differentiated rocks (Emmons, 1927, p. 75–76; Cornwall, 1951, p. 162–163; Hotz, 1953, p. 683) and may reflect the ascent of late magmatic fluids. The pattern of alteration is therefore deuteric, but the high water content (average 3.18 per cent) suggests further alteration. Probably this is due to the effect of successively later gabbro intrusions, each of which would drive water from sediments into earlier gabbros. In the basalts alteration is slight.

The significance of pumpellyite and prehnite in some of the gabbros needs consideration in view of their role in the zeolite facies (Coombs, 1954; Fyfe *et al.*, 1958, p. 215–217). Coombs (1954) found that Triassic tuffs and graywackes in Southland, New Zealand, buried to depths of 20,000–30,000 feet, are completely recrystallized to an assemblage, laumontite-albite-quartz± prehnite, calcite, and epidote. One particular bed low in the sequence has quartz-adularia-albite-pumpellyite, formed by concentration of lime and alumina released during albitization of plagioclase. Turner (*in* Fyfe *et al.*, 1958, p. 216) considers that these assemblages meet the definition of metamorphic facies and takes them as a basis for the zeolite facies. Conditions leading to the formation of the New Zealand zeolite facies are: (1) a sedimentary assemblage rich in highly reactive volcanic glass; (2) initial enrichment of the assemblage in soda by reaction of the glass with sea water to produce analcite. The zeolite facies cannot be expected in most metamorphic sequences, but possibly at corresponding temperatures and pressures some of the minerals develop where local compositions are favorable.

In gabbros west of Murdock and Irene lakes, pumpellyite and prehnite may be found in plagioclase or in interstices, particularly in albite gabbro pegmatites. Since the gabbros are on the low-temperature side of the quartz-albite-muscovite-chlorite subfacies, pressure-temperature conditions should approach those of the zeolite facies, and development of pumpellyite and prehnite is compatible with the metamorphic environment. Indeterminate mesostasis in albite-gabbro pegmatites may be devitrified glass. It is tempting to regard the albite of albite gabbro pegmatite together with quartz, pumpellyite, and prehnite as partial development of the zeolite facies. Further discussion is deferred until magmatic aspects of albite development have been considered.

Contact Metamorphism

Contact metamorphism in sedimentary rocks west of Murdock-Irene lakes is surprisingly slight. The principal effect appears to be recrystallization to a finer grain size. Rounded quartz grains, 0.1–0.2 mm in diameter, are recrystallized to a fine-grained mosaic of interlocking crystals. Chlorite is generally present in minute flakes parallel to bedding. Biotite is minor in contact specimens but absent in specimens farther from the contact; hence it may be a product of contact metamorphism. It is generally fine-grained and lacks the usual appearance of a metamorphic mineral. Only two specimens showed incipient development of porphyroblasts. Rounded clots up to 0.1 mm in diameter are scattered through a chloritic shale. Each clot is a mass of chlorite flakes with common orientation. Some have clear centers. The host mineral could not be identified, but its appearance suggests cordierite.

Slight contact metamorphic effects produced by mafic sheets are not uncommon (Geikie, 1896, p. 376; Harker, 1904, p. 245; Grout and Schwartz, 1933, p. 25) but are surprising in view of the high temperatures of basaltic magmas. Jaeger (1957) showed that the contact temperature of a dolerite sill with a range of solidification of 1100°–800°C. would be 665°C. If the sill were 200 m thick a temperature above 600°C. would be maintained for 684 years. Country rock 20 m from the intrusion would remain at a temperature higher than 550°C. for 480 years (Jaeger, 1957, p. 315). Thus it is difficult to account for the apparent lack of high-temperature thermal metamorphism. Recrystallization of rocks to finer grain size, suggestive of water-deficient metamorphism(Poldervaart, 1953, p. 265), may be sufficient manifestation of high contact temperatures under these conditions.

Igneous Petrology

Normal Gabbro Series

Included in the normal gabbro series are normal gabbros, basalts, and metagabbros. The relationship of normal gabbros and basalts is shown by the smooth variation curves and the characteristic chemistry marked by unusually low potassium and strontium. Metagabbros on the basis of distribution and texture appear to be metamorphic equivalents of the normal gabbros, but their chemical relationship to the normal gabbros cannot be demonstrated with certainty. Leopard rock and leuco- and melanometagabbro are considered separately.

PARENT MAGMA: Since analyses of the chilled phases proved unreliable, the composition of the parent magma was found indirectly. Rocks at the mafic end of the series which contain abundant olivine are not on the liquid line of descent. They are near the base of sills and represent a small proportion of the total volume of gabbro and basalt. The first rock of the series known to be on the liquid line of descent (*see* Figure 16) is the lava specimen B54A. This is followed by the composite sample of Doublet basalt (Fahrig, 1953, Ph.D. thesis, Univ. Chicago, p. 8) and the subpegmatitic gabbro (R129). The lavas must be near the beginning of the liquid line of descent and should approach the parent magma. The true average of a differentiated series should be the same as the parent magma. Without knowing the proportions represented by each analysis it is impossible to compute the true average, but the arithmetic average is probably a close approximation. The average of the normal gabbro series (excluding metagabbros) and the basalt analyses are given in Table 9. The three analyses are similar, especially the average normal gabbro and basalt B54A; hence the parent magma is assumed to be represented by analysis B54A. Differentiation must have taken place entirely *in situ*.

DIFFERENTIATION PROCESS: Differentiation in the Ahr Lake gabbros apparently took place by crystal fractionation. In view of the recent discussion on differentiation (Walker, 1956;

Jaeger and Joplin, 1956; Hess, 1956), the mechanism involved requires careful consideration.

Jaeger and Joplid (1955, p. 13–17) challenge the concept of magmatic differentiation through crystal settling. They argue that crystals formed near the upper margin of a sill cannot survive long enough in the hotter magma below to settle through the sill to the base. In place of settling by individual crystals they suggest settling of solid blocks stoped from the roof or of blocks of crystal mush torn from a solidifying layer near the roof. Such a layer forms when crystallization moves downward faster than crystals can settle. Walker (1956, p. 434) claims that most dolerite sills started to crystallize before intrusion, as shown by phenocrysts in the chilled margins. The hypothesis of Jaeger and Joplin lacks field support, although it should produce distinctive petrographic features (Walker, 1956, p. 436–439). Jaeger and Joplin (1956, p. 443) reply that the presence of phenocrysts in the chilled margins is not proof of slight super-cooling. Hess (1956, p. 446) adds that microphenocrysts in the border facies of the Palisade sill show evidence of resorption, indicating that the magma was superheated. The presence of unfused arkose xenoliths in the Palisades sill limits superheat to less than 50°C. (Hess, 1956, p. 447). Hess (1956, p. 449) contrasts slow settling of individual crystals with more rapid settling in convective overturns and concludes that convective circulation is almost inevitable. Igneous lamination results from crystal settling provided that crystals are not equidimensional. Dolerite sills do not have igneous lamination, and Hess (1956, p. 450) suggests an alternative differentiation process which involves crystallization inward from cooling surfaces, with the magma at the point of crystallization continually changing. Continuation of this process results in a highly differentiated sill. To be expected from this process is a similar but asymmetrical differentiation inward from roof and floor. Hess (1956, p. 450) demonstrated this for the Lambertville sill, where the differentiated roof zone is about a quarter of the thickness of the floor zone.

TABLE 9.—AVERAGE NORMAL GABBRO AND BASALT ANALYSES
(Recalculated water-free, to 100 per cent)

	1	2	3
SiO_2	50.07	49.74	51.84
TiO_2	1.32	1.56	0.55
Al_2O_3	14.79	14.47	15.68
Fe_2O_3	1.44	1.90	2.48
FeO	11.69	12.45	9.99
MnO	0.22	0.25	0.38
MgO	7.55	6.86	6.57
CaO	10.52	10.20	10.00
Na_2O	2.06	2.20	1.82
K_2O	0.25	0.22	0.45
P_2O_5	0.09	0.15	0.24
	100.00	100.00	100.00

1. Basalt B54A, west of Murdock Lake
2. Average normal gabbro (14 analyses)
3. Composite sample of Doublet metabasalt, Griffis Lake area (Fahrig, 1953, Ph.D. Thesis, Univ. Chicago, p. 8)

The following points are pertinent to the problem of differentiation in the Ahr Lake sills: (1) Chilled margins contain microphenocrysts of plagioclase, olivine pseudomorphs, and augite. (2) Equidimensional plagioclase phenocrysts, 2–6 mm in diameter, in the lower portion of sills are commonly euhedral. Locally severe saussuritization makes crystal outlines indistinct, but resorption textures seem to be absent. (3) Igneous lamination is not apparent. (4) No structures were observed which might be interpreted as settled blocks.

Plagioclase phenocrysts at the base of sills are probably intratelluric and settled after the sills were emplaced. Otherwise Ahr Lake sills show little clear evidence of their method of differentiation. A differentiated zone adjacent to upper margins of sills has not been found, but sampling near upper contacts was poor because of paucity of exposures on dip slopes, and narrow upper zones may have been missed. Probably both crystal settling and marginal differentiation were operative, but it is not possible to state which predominated.

TREND OF DIFFERENTIATION: The iron-enriched trend of the normal gabbro series is undoubtedly a consequence of water deficiency. Kennedy (1955, p. 500–502) states that in "wet" magma iron is oxidized to a greater degree than in "dry" magma because of free oxygen available from dissociation of water. Iron crystallizes early as a spinel, and the liquid changes in composition toward diorite. In a water-deficient magma iron remains predominantly as ferrous iron and is concentrated in the liquid phase; hence the magma shows

iron enrichment. These views are confirmed by work on the system MgO-FeO-Fe_2O_3-SiO_2 (Osborn, 1958, p. 1626). Osborn suggests that basaltic magmas of orogenic belts contain enough water to differentiate along the calc-alkaline trend, whereas basaltic magma of sills and extrusives are poor in water and follow a trend of iron enrichment. Few basaltic provinces show as extreme a trend as the Ahr Lake gabbros (*c.f.*, Walker and Poldervaart, 1949, p. 657–661); hence the Ahr Lake magma must have been exceptionally "dry." Nevertheless, most of the sills contained sufficient water to produce coarse-grained pegmatitic phases in their upper portions.

The present high-water content of the gabbros (Table 11) is attributed to the action of continued sill intrusion. Each succeeding intrusion would drive water from sediments into previously consolidated sills causing extensive alteration.

CONTRASTING DIFFERENTIATION PRODUCTS: Differentiation in Ahr Lake sills produced two end products of distinctive texture and mineralogy: gabbro pegmatite and albite gabbro pegmatite. The series ending with gabbro pegmatite is typical and represented in the idealized sill (Fig.5). Albite gabbro pegmatite is found in one sill only, and stages leading to its development are incompletely known. It overlies highly altered, even-grained gabbro and in one place grades into a transitional phase. The two series can be represented schematically as follows:

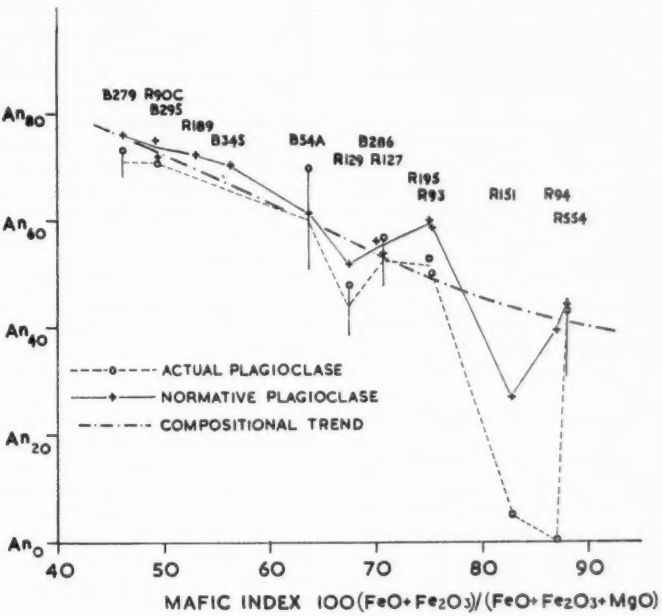
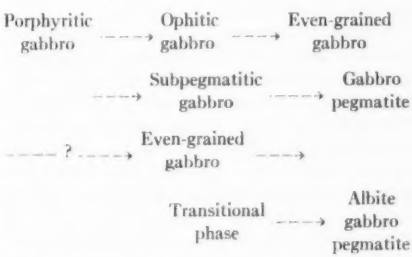


FIGURE 18.—VARIATION OF NORMATIVE AND ACTUAL PLAGIOCLASE WITH MAFIC INDEX



Both series probably developed from the same parent magma. Features characterizing the province as a whole are common to each series: the differentiation trend toward extreme iron enrichment and unusually low strontium and potassium. In the variation diagram (Fig.16)

neither series shows consistent departures from the average trends, except silica. The mineralogy of albite gabbro pegmatite and gabbro pegmatite is approximately equivalent (Fig.18). Normative plagioclase of albite gabbro pegmatite is not greatly different from the trend of normative plagioclase in the entire series. Calcium in plagioclase of gabbro pegmatite must be represented by other calcium minerals in albite gabbro pegmatite.

ORIGIN OF ALBITE GABBRO PEGMATITE: The rock is characterized by elongated plagioclase, thin plates of ilmenite, and a finely crystalline mesostasis which may be devitrified glass. Locally a fine-textured micropegmatite is present. Walker (1956, p. 440–441) refers to pegmatities in some Tasmanian dolerite sills which carry devitrified glass in place of micropegmatite. These may include the iron-rich portions of the Great Lake dolerite described by Jaeger and Joplin (1955, p. 10–12), which contain large areas of mesostasis. Of the conditions likely to lead to glass formation, escape of volatiles seems most applicable in the present case. Albite gabbro pegmatites formed from water-deficient magma judging from their high FeO-Fe_2O_3 ratios (Kennedy, 1955, p. 496) and differentiation trend (Osborn, 1958, p. 1626), but this is common to both series. Gabbro pegmatite is coarser than albite gabbro pegmatite, which suggests that it was richer in volatiles. Perhaps minor differences in volatile content are sufficient to cause glass to form in one rock and micropegmatite in another, but this would have negligible effects on differentiation trends and iron ratios. The sill containing albite gabbro pegmatite may have lost volatiles during emplacement or crystallization and solidified in the final stages to glass.

The albite may be magmatic or metamorphic in origin. In several other localities where albite is found in little metamorphosed mafic rocks it has been interpreted as magmatic. Patches in the Höllefors dolerite and porphyrite (Krokström, 1937) contain albite, whereas plagioclase in surrounding rocks is andesine. Krokström (1937, p. 206) postulated that segregations of volatiles in magma extended the range of crystallization locally to lower temperatures. As a result early formed andesine was albitized, and excess calcium formed epidote. Pegmatites in Keweenawan lavas contain albite-oligoclase, whereas the surrounding ophitic basalt has labradorite or andesine. The pegmatitic plagioclase is altered to prehnite, pumpellyite, chlorite, epidote, sericite, and kaolinite (Cornwall, 1951, p. 162). Niggli (1952, p. 406) suggested that with enrichment of pegmatitic zones in water, calcic plagioclase is replaced by hydrous calcium minerals and albite. Battey (1956, p. 103) believes that albite in spilites of North Island, New Zealand, formed under hydrous magmatic conditions. Battey postulated that plagioclase and augite crystallized together, but in a low-temperature, hydrous environment calcium is eliminated from plagioclase and is partly concentrated in the liquid and partly fixed in augite. Iron is concentrated in the liquid because of failure of pigeonite to crystallize. Calcium and iron are then used in hydrous minerals such as babingtonite, epidote, pumpellyite, prehnite, and iron-rich chlorite.

According to these authors the presence of albite in mafic rocks depends on hydrous conditions. Volatiles may lower crystallization temperatures of magma below the stability of calcic plagioclase and provide an environment which promotes diffusion and reaction. Late fractions of initially water-deficient magmas may approach similar temperatures, evidenced in many mafic pegmatites by interstitial alkali feldspar and chlorite, but paucity of volatiles probably slows reactions with early formed minerals which persist at the lower temperatures. If albite of albite gabbro pegmatite is magmatic, the hydrous conditions necessary for its development are difficult to reconcile with indications of water deficiency given by iron ratios and the differentiation trend. If the mesostasis is devitrified glass, denoting a viscous end magma, the case for primary albite is further weakened. Gabbro pegmatite gives every indication of having formed from a more hydrous magma than albite gabbro pegmatite yet contains a more calcic plagioclase.

The albite may be considered metamorphic, and with pumpellyite, prehnite, and quartz it may represent partial development of the zeolite facies. At low metamorphic temperatures, reactions may be expected in rocks that have a glass phase but would be to slow to affect holocrystalline rocks. This would account for the limited distribution of albite. However, many specimens of albite gabbro pegmatite contain neither pumpellyite nor prehnite, and if albite is metamorphic, calcium must have been removed. Pumpellyite and prehnite are found locally in gabbro pegmatite, subpegmatitic gabbro, and rarely in more mafic varieties of gabbro, where plagioclase is not albite. They may represent an incipient stage of metamorphism which was prevented from

more nearly complete development by low rates of diffusion and reaction.

The writer concludes tentatively that albite in albite gabbro pegmatite is metamorphic and with pumpellyite, prehnite, and quartz may indicate the zeolite facies.

SEQUENCE OF EVENTS: The writer postulates the following sequence of events:

(1) Sedimentary and volcanic rocks in the eastern part of the Labrador trough were profusely invaded by magma of essentially uniform composition (excluding leopard rocks).

(2) A few plagioclase and possibly olivine and pyroxene phenocrysts carried in by the magma settled to the bottom when movements ceased.

(3) Differentiation continued as the sills crystallized, probably partly by crystal fractionation at sill boundaries (Hess, 1956, p. 450) and partly by convective overturn and crystal settling.

(4) The magma was unusually low in water; consequently iron persisted in the reduced state and entered silicates in progressively greater amounts as differentiation proceeded. Differentiation was predominantly toward iron enrichment, and an alkali-rich liquid failed to separate in the late stages.

(5) Most sills contained sufficient water to end crystallization with coarse-grained pegmatites similar to gabbro pegmatities elsewhere (Walker, 1953). One sill with slightly less water completed its solidification with interstitial glass.

(6) Regional metamorphism west of Murdock and Irene lakes was too mild to have extensive effects on the rocks. Locally pumpellyite and prehnite developed in gabbro where conditions were favorable. Gabbro with a glass phase was probably more susceptible to metamorphism and locally approached zeolite facies.

Leopard Rock

Sills of leopard rock are associates of gabbro sills throughout the Labrador trough. Under the names blotchy gabbro, mottled gabbro, or leopard rock they have been described by Frarey (1952, p. 5–6) in the Willbob Lake area, 30–50 miles southeast of the Ahr Lake area; Fahrig (1955) in the Lac Herodier area, just south of Koksoak River; and by Bergeron (1954, p. 4–5; 1956, p. 4), Sauvé (1955, p. 4; 1956, p. 5), and Bréard (1957, p. 6) in the region north of Koksoak River. They are therefore a characteristic feature of the Labrador basaltic province.

Without later intrusions of gabbro magma, leopard-rock sills consist of a main zone of leopard rock flanked by thin marginal zones of gabbro with chilled contacts.

TABLE 10.—ANALYSES OF VARIETIES OF LEOPARD ROCK AND PARENT BASALT
(Recalculated water-free, to 100 per cent)

	B54A	R161	R163	R177	181
SiO_2	50.07	50.28	50.43	48.96	46.90
TiO_2	1.32	1.65	1.61	0.94	0.27
Al_2O_3	14.79	15.50	16.95	21.25	29.90
Fe_2O_3	1.44	0.88	0.82	1.37	2.13
FeO	11.69	12.80	13.09	7.70	0.76
MnO	0.22	0.14	0.19	0.15	0.03
MgO	7.55	7.48	3.93	4.22	1.42
CaO	10.52	6.12	10.09	11.74	14.60
Na_2O	2.06	2.17	2.32	2.94	2.70
K_2O	0.25	2.74	0.48	0.59	1.26
Rest	0.09	0.24	0.09	0.14	0.03

B54A. Basalt, presumed parent of normal gabbro series

R161. Chilled contact, leopard-rock sill

R163. Border zone gabbro, leopard-rock sill 2 feet from contact

R177. Leopard rock

R181. Feldspathic clot in leopard rock

RELATIONSHIP OF LEOPARD ROCK TO NORMAL GABBRO SERIES: Chemical analyses from various phases of a leopard-rock sill together with the basalt analysis (B54A) are given in Table 10. If the effects of metasomatism (high potassium, low calcium) are neglected, the chilled contact is similar to the basalt regarded as parent magma of the normal gabbros. A genetic relationship between normal gabbros and the border phase of leopard rock is also indicated by the low strontium content. The border phase passes into leopard rock by a marked increase in number and size of feldspathic clots. Thus leopard rock must be derived from a magma similar to the parent of the normal gabbro series by addition of plagioclase.

EMPLACEMENT: Presumably leopard rock magma was intruded with a load of crystals and crystal aggregates. Recent work on the flow of pulp suspensions done for the paper industry (Robertson and Mason, 1956; Baines, 1956) may explain why there are no crystal aggregates in the chilled margins and why they are scarce in the border zone. The

experiments were made on suspensions of wood fibre, but the broad outlines of flow structure should apply to any granular particles in a medium of approximately the same specific gravity. Baines (1956, p. 40) recognized four regimes in the flow of a dilute suspension of sulfite pulp in a smooth pipe.[6] At low velocities pulp settles to the bottom and moves with internal shearing. As velocity is increased past a critical value, all the fiber becomes a central plug which moves at a uniform velocity. This is surrounded by an annulus of water in laminar motion. With further velocity increases the flow passes through two more critical values, first to a regime of a plug surrounded by turbulent flow, then to fully turbulent flow.

The plug flow described by Baines fits the picture of a leopard-rock sill with a uniform core of feldspathic aggregates and marginal gabbro zones. In a sill the "plug" would be tabular; hence flow is better described as sheet flow. When movements ceased the magma was probably sufficiently viscous to preserve the divisions of flow structure. The laminar structure at the boundary between leopard rock and border phase at one place (Pl.3, fig.2) indicates that laminar flow existed in the border region. Elsewhere, the structure may have been obscured by slight settling of particles after flow had stopped.

ORIGIN: Four possible explanations for plagioclase concentrations in leopard rock are: (1) Plagioclase was torn from pre-existing anorthositic masses at depth. Grout and Schwartz (1939, p. 57–62) suggested this explanation for the origin of anorthositic masses in portions of the Logan sills, Minnesota. These are rounded to angular masses which range from a single crystal to a quarter of a mile across and are in places faintly banded. Photographs (Grout and Schwartz, 1939, Fig.36, p. 51) show dolerite thickly packed with angular fragments. In contrast, leopard rock contains regularly spaced, rounded feldspathic masses in a groundmass of fairly uniform grain size. Angular fragments or truncated crystals are missing. If the feldspar was torn from anorthositic masses it must have been considerably modified en route from the source.

(2) Plagioclose crystallized from normal gabbro magma and was concentrated before final emplacement. The formation of leopard rock requires plagioclase concentration without concentration of ferromagnesian minerals. Plagioclase concentrated at the bottom of a magma reservoir by crystal settling is likely to be associated with ferromagnesian accumulates and is also unfavorably located for reintrusion. Plagioclase suspended in a magma reservoir[7] might be concentrated by eddy currents in hollows or magma projections in wall rock. Heavier ferromagnesian crystals would tend to settle and would be less subject to current movements; hence they would not concentrate with plagioclase. The slight difference in specific gravity between labradorite—2.68 at 1100°C. (Kosu and Ueda, 1933, p. 262)—and basaltic magma—2.60-2.64 (Dane, 1942, p. 36)—makes suspension of plagioclase feasible. Plagioclase concentrated in the upper parts of a magma reservoir would be in a favorable position for early expulsion and would be followed by magma with few plagioclase phenocrysts. This may explain the observed sequence: leopard rock-sparsely porphyritic gabbro.

(3) Plagioclase was derived from the earth's mantle by selective melting. The difference in melting temperatures between albite and anorthite decreases with increasing pressure until at approximately 22,000 bars they melt at the same temperature (Yoder, 1952, p. 373). At greater pressures anorthite presumably melts first. Verhoogen (1954, p. 87–88) has suggested that high-melting-point minerals at the surface may be the low-melting-point minerals at depths of 100 km or greater. Partial fusion at the proper depths may lead to separation of a liquid with an unusually high proportion of intermediate plagioclase. If crystallization in such a liquid began en route to the surface or during a pause at intermediate depths, the final intrusion should be rich in plagioclase crystals. Partial melting may result from release of pressure (Yoder, 1952, p. 371). With the initial release of pressure anorthositic liquids may develop, but subsequently pressures may be lowered to the point where normal basaltic magmas are produced. This may account for the initial intrusion of leopard rock, followed by normal gabbro magma. Similarities in composition between the border phase of leopard rock and the normal gabbro series are to be expected since they originate in the same part of the mantle.

[6] The flow profile should be similar in a smooth-walled sill.

[7] Holmes and Harwood (1929, p. 44) account for anorthite aggregates in tholeiitic dikes of the north of England by a process involving suspension of early formed anorthite.

(4) Gabbroic anorthositic magma may be the water-rich equivalant of normal gabbroic magmas (Buddington, 1956, p. 100). According to Yoder's data (1955, p. 106–107) on the system water-anorthite-diopside the eutectic ratio for anorthite diopside is shifted from 42 per cent anorthite in the dry system to 73 per cent anorthite in a system with 9 per cent H_2O at 5000 atmospheres pressure and 1095°C. Buddington reasons that the cotectic ratio of labradorite to mafic minerals would be even higher than for anorthite and may approach 85 per cent feldspar, the actual ratio in gabbroic anorthositic magmas of the Adirondacks. The leopard-rock ratio of feldspar to mafic minerals of approximately 70 to 30 (normative) should be readily obtained with partial melting of the mantle under conditions of moderately high water pressure. The water may have been lost following the first expulsions of magma, so that later intrusions were of normal basalt.

The large quantity of plagioclase-rich magma required for widespread occurences of leopard rock in the Labrador trough favors the explanations of selective melting. Concentration of plagioclase by currents in a magma reservoir should produce local enrichments only.

Regardless of how plagioclase was initally concentrated, it would probably be further concentrated during the regime of sheet flow. Part of the liquid phase at the margins would form chilled contacts and be removed from the system, and liquid would be drawn from the central sheet to replace it. Probably the proportion of liquid phase decreases, according to how far the sill is from its source, until the flow ceases because of lack of lubricant. Further movements would have to be along slip planes. This has not been tested in the field.

SURFACE FLOWS: Flows and associated intrusive rocks that contain sparsely scattered clots of feldspar about the size of golf balls have been described. In two places the associated gabbro is unusually coarse-grained and contains abnormally large feldspar clusters. These rocks are similar to leopard rock (*c.f.*, Figure 1 of Plate 2; Figure 1 of Plate 3) and seem to be surface or near-surface manifestations of leopard rock. The paucity of phenocrysts in the rocks may be related to their presence at the surface. Probably they had lower viscosity than leopard rock and could reach the surface more readily. Presence in the Doublet group of three levels of blotchy flows and near-surface intrusions may indicate that generation of leopard rock is cyclic, but repetition of one layer of blotchy flows by faulting cannot be excluded.

Leucometagabbro and Melanometagabbro

Sills grading up from melanometagabbro to leucometagabbro are not due to normal differentiation for the following reasons: (1) despite differences in color, melanometagabbro has only a slightly lower mafic index (40) than leucometagabbro (46); (2) normative plagioclase is nearly the same in both rocks (An_{68} for leucometagabbro, An_{61} for melanometagabbro); (3) leucometagabbro is only slightly enriched in alkalies and silica relative to melanometagabbro; (4) trace elements such as chromium and vanadium do not show the behavior characteristic of normal differentiation series.

Both rocks are obviously mafic, but one contains a preponderance of the lighter constituents and the other the heavier constituents of mafic rocks. In effect, the rocks are mineralogical supplements of one another. It is tempting to interpret the division of these sills into heavier and lighter fractions as due to gravity separation. However, even if the original magma were dense enough to float plagioclase, a nearly complete separation probably could not be obtained by simple floating and sinking. Possibly, intrusion of a mafic magma, 50–60 per cent crystallized, would produce gravity separation by action analogous to that of a jig used in commercial mineral separation. Jostling of crystals against one another during magma movements would allow heavier crystals to work to the bottom, displacing lighter crystals upward. The liquid medium need not be heavier than the plagioclase crystals. Either heavy or light fractions may intrude beyond the termination of the other and produce apparently independent sills.

Such a method of origin is highly spectulative. The only positive conclusion is that leucometagabbro and melanometagabbro are not normal differentiation products.

MAGMA TYPE

Introductory Notes

Basaltic magmas may give rise to contrasting differentiation suites; therefore a classification that includes both parent magma and its products is necessary. The most widely used is the classification of Kennedy (1933) who divides basalts into tholeiites and olivine basalts. Tholeiites give rise to quartz-bearing

1634, Table 11, under Composite metabasalt, omit 3.02; specific gravity under Metagabbros, *for* sample B 171, *read* sample B 71

TABLE 11.—CHEMICAL ANALYSES AND NORMS OF LABRADOR ROCKS

Index	Normal gabbros													Basalt		Leopard rock				Metagabbros					
	B279	B295	R90C	R189	B345	R129	B286	R127	R195	R93	R151	R94	R554	B54A	Composite metabasalt	R161	R 163	R 177	R 181	R 337	R 293	46–50 6–50	B 171	121–50	202–50 159–50
SiO_2	46.06	45.67	49.12	44.50	46.95	48.99	49.97	50.10	44.48	46.73	48.82	50.12	54.02	48.52	49.86	48.03	48.92	47.43	45.71	44.88	46.77	50.60	52.61	49.20	46.90
TiO_2	0.58	0.73	0.75	0.77	0.80	1.39	0.76	1.31	2.81	2.86	2.33	2.59	2.08	1.28	0.53	1.58	1.56	0.91	0.25	0.67	0.38	0.28	0.93	0.74	0.84
Al_2O_3	17.86	16.68	13.77	18.90	17.92	12.42	13.14	12.83	11.44	12.05	12.25	10.83	11.15	14.36	15.07	14.87	16.35	20.57	29.04	13.73	23.09	9.78	13.57	14.15	18.19
Fe_2O_3	1.29	1.40	1.17	1.26	1.48	2.61	2.00	2.07	1.19	2.25	4.01	1.96	1.44	1.40	2.39	0.84	0.81	1.33	2.11	1.10	1.42	1.08	0.98	5.12	3.58
FeO	7.81	8.55	8.33	7.74	7.77	12.74	11.42	12.17	17.46	14.66	13.88	18.16	16.19	11.34	9.63	12.27	12.60	7.44	0.74	8.61	4.24	9.68	7.84	10.95	7.71
MnO	0.17	0.18	0.20	0.18	0.20	0.25	0.26	0.23	0.26	0.28	0.26	0.24	0.36	0.22	0.37	0.14	0.19	0.15	0.03	0.17	0.07	0.12	0.21	0.09	0.07
MgO	10.58	10.24	9.71	7.99	7.15	7.41	5.74	5.93	6.16	5.57	3.76	2.97	2.36	7.33	6.34	7.15	3.82	4.08	1.38	14.34	6.64	11.81	8.82	4.99	6.20
CaO	10.68	10.67	13.18	12.17	12.15	8.55	10.94	10.75	10.37	10.39	7.19	4.26	6.59	10.21	9.64	5.88	9.75	11.32	14.22	8.66	12.55	11.73	7.80	8.73	11.23
Na_2O	1.37	1.44	1.34	1.80	1.86	2.34	2.19	2.30	1.71	1.83	4.26	2.76	2.57	2.05	1.76	2.05	2.26	2.80	2.63	1.95	2.66	1.88	3.90	2.56	1.60
K_2O	0.23	0.10	0.05	0.26	0.32	0.12	0.28	0.12	0.31	0.48	0.01	0.19	0.18	0.24	0.43	2.62	0.46	0.57	1.23	0.06	0.10	0.34	0.14	0.75	0.53
P_2O_5	0.23	0.05	0.05	0.03	0.12	0.00	0.11	0.00	0.06	0.30	0.10	0.59	0.25	0.09	0.20	0.21	0.08	0.10	0.03	0.25	tr.	0.08	0.09	0.24	0.23
Cr_2O_3	0.02	tr.	0.13	0.16	tr.	n.d.	tr.	n.d.	tr.	n.d.	n.d.	tr.	n.d.	tr.	tr.	0.02	n.d.	tr.	n.d.	0.07	n.d.	0.09	0.03	n.d.	n.d.
H_2O^+	3.26	3.99	2.20	3.76	3.09	3.73	2.82	2.57	3.03	2.62	3.34	4.08	3.02	2.88	3.18	4.02	2.63	2.85	2.56	4.91	1.92	1.89	2.62	1.46	2.24
H_2O^-	0.25	0.36	0.16	0.25	0.20	0.09	0.12	0.09	0.24	0.11	0.43	0.35	0.00	0.21	0.34	0.27	0.08	0.19	0.17	0.11	0.00	0.11	0.13	0.31	0.18
CO_2	0.01	0.08	0.05	0.01	0.01	0.00	0.27	0.00	0.04	0.00	n.d.	0.22	0.00	0.00	tr.	0.00	0.05	0.00	0.00	0.09	n.d.	0.00	0.00	0.36	0.27
S	0.01	0.07	0.06	0.02	0.03	n.d.	0.15	n.d.	0.18	n.d.	n.d.	0.17	n.d.	0.06	0.15	0.03	n.d.	0.03	n.d.	0.06	n.d.	0.10	0.00	0.18	0.16
Total	100.40	100.21	100.27	99.70	100.05	100.64	100.17	100.47	99.74	100.13	100.64	99.65	100.21	100.19	99.89	99.98	99.56	99.77	100.10	99.66	99.84	99.57	99.67	99.91	99.93
Less O≡S	0.01	0.03	0.03	0.01	0.01	..	0.06	..	0.08	..	..	0.07	..	0.03	0.06	0.01	..	0.01	..	0.03	..	0.04	..	0.07	0.06
Net Total	100.39	100.18	100.24	99.69	100.04	100.64	100.11	100.47	99.66	100.13	100.64	99.58	100.21	100.16	99.83	99.97	99.56	99.76	100.10	99.63	99.84	99.53	99.67	99.84	99.87
Sp. gr.	2.98	3.01	3.06	2.99	3.02	2.98	3.04	3.02	3.09	3.05	2.99	3.08	2.97	3.02	3.02	..	..	2.99	..	2.95	2.96	..	2.94	..	..
Mafic Index	46.2	49.3	49.4	53.0	56.4	67.5	70.0	70.7	75.0	75.2	82.7	87.0	88.0	63.7	65.5	64.7	76.8	68.3	..	40.3	46.0	47.7	50.0	76.3	64.5

Index	Normal gabbros													Basalt		Leopard rock				Metagabbros					
	B279	B295	R90C	R189	B345	R129	B286	R127	R195	R93	R151	R94	R554	B54A	Composite metabasalt	R 161	R163	R177	R181	R337	R293	46–50 6–50	B71	121–50	202–50 159–50
Qu	..	..	..	..	..	0.90	2.66	2.05	..	0.73	..	8.64	12.46	..	4.64	..	1.73	..	..	..	..	..	..	3.50	1.80
Or	1.34	0.61	0.28	1.56	1.89	0.72	1.67	0.72	1.84	2.84	0.06	1.11	1.06	1.39	2.56	15.48	2.73	3.40	7.29	0.33	0.61	2.00	0.84	4.45	3.12
Ab	11.59	12.17	11.33	15.21	15.74	19.77	18.52	19.46	14.48	15.47	36.03	23.34	21.77	17.36	14.90	17.36	19.14	23.71	11.28	16.52	20.09	15.89	32.99	21.66	13.53
An	41.93	38.76	31.44	42.74	39.62	23.04	25.21	24.32	22.62	23.26	14.27	16.61	18.36	29.27	31.94	23.62	33.11	41.90	63.80	28.55	50.78	17.25	19.11	24.90	40.90
Ne	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	5.94	..	1.31	..	..	..	..
Di Wo	3.97	5.56	13.89	7.27	8.33	8.10	11.12	12.12	11.83	11.00	8.67	1.43	5.29	8.70	6.09	1.74	5.94	5.69	2.75	5.09	4.80	16.87	7.95	6.08	4.81
En	2.52	3.43	8.48	4.30	4.75	3.90	4.88	5.31	4.41	4.44	2.96	0.32	1.05	4.32	3.03	0.82	1.96	2.67	2.38	3.39	3.23	10.21	4.80	2.78	2.82
Fs	1.20	1.81	4.63	2.61	3.22	4.09	6.21	6.78	7.64	6.66	5.95	1.20	4.62	4.21	2.93	0.90	4.17	2.96	..	1.32	1.20	5.74	2.72	3.25	1.76
Hy Of En	12.95	12.29	15.45	0.49	7.20	14.57	9.42	9.46	5.42	9.43	5.35	7.08	4.82	12.97	12.75	8.71	7.55	1.47	..	8.18	..	12.35	13.00	9.65	12.62
Fo	6.19	6.48	8.46	0.30	4.87	15.32	12.02	12.10	9.37	14.18	10.73	26.32	21.15	12.67	12.28	9.50	16.07	1.64	..	3.19	..	6.94	7.35	11.22	7.86
Ol Fa	7.62	6.85	0.17	10.59	4.10	..	..	..	3.86	..	0.74	..	..	0.68	..	5.80	..	4.22	0.74	16.91	9.32	4.80	2.92	..	..
Il	4.00	3.99	0.10	7.12	3.05	..	..	..	7.37	..	1.64	..	..	0.73	..	6.97	..	5.16	..	7.25	3.80	2.98	1.82	..	..
Mt	1.11	1.38	1.43	1.46	1.52	2.64	1.44	2.49	5.34	5.43	4.43	4.92	3.97	2.43	1.00	3.01	2.96	1.73	0.49	1.28	0.71	0.53	1.76	1.40	1.59
Ap	1.88	2.04	1.69	1.83	2.15	3.77	2.89	3.01	1.74	3.27	5.81	2.85	2.08	2.04	3.47	1.23	1.18	1.92	1.71	1.60	2.06	1.58	1.41	7.43	5.19
Pyr	0.53	0.13	0.13	0.07	0.26	..	0.26	..	0.13	0.69	0.23	1.38	0.59	0.20	0.46	0.49	0.26	0.23	0.07	0.59	..	0.20	0.20	0.56	0.53
Calc	0.04	0.20	0.14	0.04	0.05	..	0.33	..	0.38	..	..	0.38	..	0.14	0.31	0.05	..	0.07	..	0.17	..	0.03	..	0.38	0.34
Chrom	0.02	0.18	0.11	0.02	0.02	..	0.61	..	0.09	..	..	0.50	..	..	..	..	0.11	..	..	0.20	..	..	..	0.82	0.61
Hem	..	..	0.20	0.09	..	..	..	..	..	..	..	..	..	..	..	0.02	..	..	..	0.11	..	0.13	0.05	..	..
H_2O	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	0.93	..	..	..	..	..	..
Total	3.51	4.35	2.36	4.01	3.29	3.82	2.94	2.66	3.27	2.73	3.77	4.43	3.02	3.09	3.52	4.29	2.71	3.04	2.73	5.02	1.92	2.00	2.75	1.77	2.42
	100.40	100.23	100.29	99.71	100.06	100.64	100.18	100.48	99.79	100.13	100.64	100.51	100.24	100.20	99.88	99.99	99.62	99.81	100.11	99.70	99.83	99.50	99.67	99.85	99.90

B279. Porphyritic gabbro. Analyst D. E. McKay, G.S.C.
B295. Ophitic (spotted) gabbro. Analyst J. A. Maxwell, G.S.C.
R90C. Ophitic gabbro. Analyst D. E. McKay, G.S.C.
R189. Ophitic gabbro. Analyst D. E. McKay, G.S.C.
B345. Ophitic gabbro. Analyst S. Courville, G.S.C.
R129. Subpegmatitic gabbro. Analyst H. B. Wiik, Helsinki, Finland
B286. Subpegmatitic gabbro. Analyst J. A. Maxwell, G.S.C.
R127. Even-grained gabbro. Analyst H. B. Wiik, Helsinki, Finland
R195. Even-grained gabbro. Analyst D. E. McKay, G.S.C.
R93. Transitional even-grained to albite gabbro pegmatite. Analyst H. B. Wiik, Helsinki, Finland
R151. Albite gabbro pegmatite. Analyst W. R. A. Baragar
R94. Albite gabbro pegmatite. Analyst D. E. McKay, G.S.C.
R554. Gabbro pegmatite. Analyst W. R. A. Baragar
B54A. Howse group basalt. Analyst J. A. Maxwell, G.S.C.
Composite metabasalt. Doublet group, Griffis Lake area. Analyst R. G. Fabry, G.S.C. (Fahrig, 1953, Ph.D. thesis, Univ. Chicago)
R161. Chilled margin, leopard-rock sill. Analyst D. E. McKay, G.S.C.
R163. Marginal gabbro, leopard-rock sill. Analyst W. R. A. Baragar
R177. Leopard rock. Analyst D. E. McKay, G.S.C.
R181. Large feldspathic clot from leopard rock. Analyst W. R. A. Baragar
R337. Melanometagabbro. Analyst J. A. Maxwell, G.S.C.
R293. Leucometagabbro. Analyst W. R. A. Baragar
46–50; 6–50. Metagabbro, Griffis Lake area. Analyst R. G. Fabry, G.S.C. (unpublished analysis loaned by W. F. Fahrig)
B71. Metagabbro. Analyst S. Courville, G.S.C.
121–50. Metagabbro, Griffis Lake area. Analyst R. G. Fabry, G.S.C. (unpublished analysis loaned by W. F. Fahrig)
202–50; 159–50. Blotchy metagabbro, Griffis Lake area. Analyst R. G. Fabry, G.S.C. (unpublished analysis loaned by W. F. Fahrig)

products, whereas olivine basalts give rise to alkaline rocks. Nockolds and Allen (1953; 1954; 1956) group their studies of igneous rock series into three subdivisions: calc-alkaline, alkaline, and tholeiitic. The calc-alkaline series includes basalts, andesites, and rhyolites, as well as batholithic masses of orogenic regions; the other two series are similar to Kennedy's. Tilley (1950, p. 55–56) suggests, in addition to tholeiites and alkali basalts, a high-alumina basaltic series exemplified by rocks of the Medicine Lake Highlands, California.

TABLE 12.—AHR LAKE MAGMA AND PARENTAL THOLEIITIC, ALKALI BASALTIC, CALC-ALKALINE, AND HIGH-ALUMINA BASALTIC MAGMAS
(Calculated water-free, to 100 per cent)

	1	2	3	4	5
SiO_2	50.07	51.5	46.6	54.0	48.10
TiO_2	1.32	1.5	2.9	0.9	0.89
Al_2O_3	14.79	14.8	15.7	18.1	18.22
Fe_2O_3	1.44	1.6	3.9	2.5	1.04
FeO	11.69	9.9	8.8	5.8	8.31
MnO	0.22	..	0.2	0.1	0.17
MgO	7.55	7.2	7.7	5.5	8.96
CaO	10.52	11.0	10.0	8.4	11.30
Na_2O	2.06	2.0	2.9	3.4	2.80
K_2O	0.25	0.5	0.9	1.1	0.14
P_2O_5	0.09	..	0.4	0.2	0.07

1. Ahr Lake basalt specimen B54A
2. Average parental tholeiites, average of 6 parental tholeiites (Nockolds and Allen, 956 p. 76)
3. Average parental alkali basalt. Average of 8 parental alkali basalts (Nockolds and Allen, 1954, p. 282)
4. Average parental calc-alkali magma. Average of 10 parental magmas (Nockolds and Allen, 1953, p. 139)
5. High-alumina basalt, Medicine Lake Highlands, California (Yoder and Tilley, 1957, p. 157)

Classification

Classification of the Ahr Lake series requires consideration of initial composition and subsequent chemistry and mineralogy. The analysis of Ahr Lake parent magma (B54A) is compared in Table 12 with averages of calc-alkaline, alkaline, and tholeiitic parental magmas (Nockolds and Allen, 1953, p. 139; 1954, p. 282; 1956, p. 76) and an analysis of high-alumina basalt (Yoder and Tilley, 1957, p. 157). The resemblance between the Ahr Lake parent magma and average parental tholeiite is striking. Parental alkali basalt has lower silica and higher alkalies, whereas the parent of the calc-alkaline series shows little similarity. Alumina in the Ahr Lake parent is too low to classify it as high-alumina basalt.

Tilley (1950, p. 42) distinguishes between tholeiites and alkali basalts of the Hawaiian Islands on a silica-alkalies variation diagram. Wager (1956, p. 231) shows that plots of Hebridean alkali basalts and tholeiites lie along discrete trends on a soda-silica diagram. In Figure 19 rocks of the Ahr Lake series are shown on an alkali-silica diagram together with the fields and trend lines of the Hawaiian and Hebridean series. The Ahr Lake trend coincides with the Hawaiian and Hebridean tholeiites.

The mineralogy of the Ahr Lake rocks is consistent with that of tholeiites. Olivine disappears early, and quartz appears in the late stages. Lime-poor pyroxene and augite are present throughout much of the crystallization period. In the alkali basalts olivine persists throughout, and augite is generally the sole pyroxene. In the late stages alkali feldspar, analcite, or a feldspathoid appears in place of quartz.

Comparison with Other Provinces

Data from many basaltic provinces have been compiled by Burri and Niggli (1945), Walker and Poldervaart (1949), Nockolds and Allen (1953; 1954; 1956), Green and Poldervaart (1955), and Wager (1956). No attempt is made here to present extensive data for comparison, but reference is made to these publications.

The bulk chemistry of the parental basalt shows little to distinguish it from other tholeiitic provinces. Three features distinguish the series: (1) a differentiation trend showing extreme iron enrichment, (2) low potassium content throughout, and (3) low strontium content throughout.

Wager (1956, p. 234–235) pointed out that rock series should be compared at equivalent stages of fractionation, to minimize differences due to fractionation alone. He proposed (1956, p. 219) a variation diagram of iron ratio ($Fe'' + Mn'' / Mg'' + Mn'' + Fe''$) against albite ratio, normative molecular $Ab/(Ab + An)$, divided into fields which cross the general

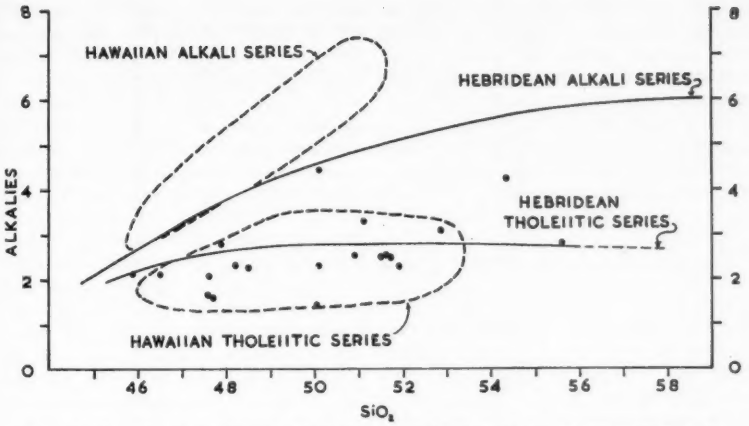


FIGURE 19.—AHR LAKE SERIES COMPARED WITH HAWAIIAN AND HEBRIDEAN SERIES
Hebridean series is based on soda silica; Hawaiian and Ahr Lake series are based on total alkalies silica

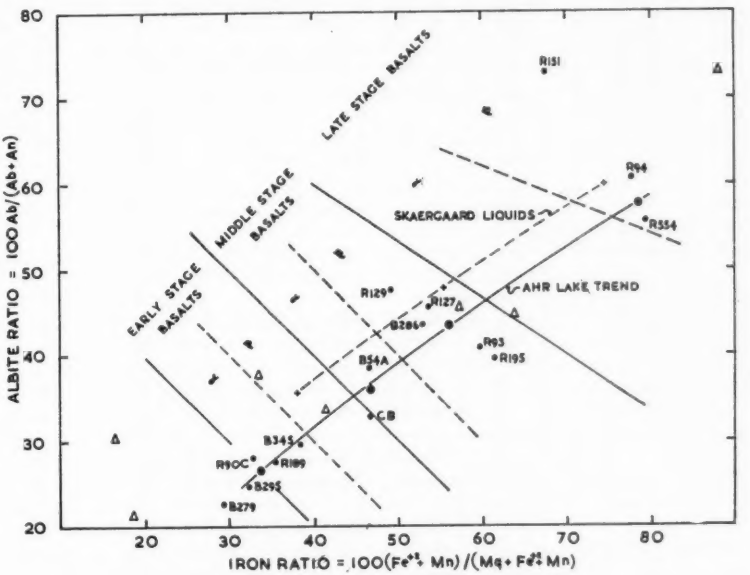


FIGURE 20.—DIFFERENTIATION DIAGRAM

Compares Ahr Lake trend with that of Skaergaard and Elephant's Head-New Amalfi intrusion. Small circles— Ahr Lake analyses; large circles— average Ahr Lake at each stage; crosses— Skaergaard liquids; triangles— Elephant's Head-New Amalfi analyses. Average early stage Ahr Lake basalt includes the primary analyses B295 and B279; average middle stage α includes the composite basalt analysis (CB); average late stage β consists of R94 and R554; R151 is abnormal.

differentiation trend and correspond to early stage, middle-stage, and late-stage differentiates. Each stage may be divided into two substages. Analyses from different series plotted on the diagram are at equivalent stages of fractionation and may be averaged and compared at each stage.

Analyses of the Ahr Lake series are plotted

TABLE 13.—AHR LAKE THOLEIITES AND OTHER THOLEIITES*
(Recalculated water-free, to 100 per cent)

	Ahr Lake					Skaergaard			Elephant's Head-New Amalfi		Hawaii			Hebrides				
	Early stage α	Middle stage		Average	Late stage β	Early stage β	Middle stage β	Late stage β	Early stage β	Middle stage β	Early stage α	Middle stage		Early stage β	Middle stage		Late stage	
		α	β									α	β		α	β	α	β
SiO_2	48.02	50.82	49.48	49.86	54.20	47.9	46.7	45.7	50.01	53.71	51.2	51.4	51.2	46.9	51.5	52.6	52.5	55.6
TiO_2	0.75	0.93	1.89	1.61	2.42	1.4	2.2	2.4	0.96	2.34	3.0	3.1	3.2	0.6	1.1	1.1	2.1	2.5
Al_2O_3	17.64	15.22	12.75	13.44	11.43	18.9	15.3	12.7	16.26	13.62	13.2	13.1	12.9	14.6	14.8	14.6	14.3	13.2
Fe_2O_3	1.38	1.96	2.08	2.05	1.77	1.2	2.9	3.6	0.32	2.26	3.6	1.7	2.3	3.3	3.1	3.2	3.3	3.5
FeO	8.33	10.85	14.10	13.17	17.88	8.7	12.9	18.2	9.30	12.20	8.3	9.5	9.5	8.8	7.5	8.1	9.1	9.2
MnO	0.20	0.30	0.27	0.28	0.31	0.1	0.2	0.2	0.11	0.23	0.1	0.1	0.1	0.2	0.2	0.2	0.2	0.5
MgO	9.46	7.07	6.34	6.55	2.76	7.8	5.9	3.5	8.84	3.74	7.6	7.5	7.2	8.5	6.8	5.7	4.5	2.9
CaO	12.19	10.88	10.51	10.43	5.64	10.5	9.9	8.3	11.95	8.60	10.4	10.7	10.4	12.0	11.3	10.1	9.2	6.8
Na_2O	1.63	1.98	2.13	2.09	2.76	2.4	2.8	3.2	1.68	2.28	2.0	2.2	2.4	2.4	2.7	2.8	3.1	2.8
K_2O	0.20	0.35	0.23	0.29	0.19	0.2	0.3	0.4	0.44	0.80	0.4	0.4	0.5	0.1	0.8	1.2	1.2	2.4
P_2O_5	0.10	0.14	0.09	0.11	0.44	0.1	0.4	0.9	0.13	0.16	0.2	0.3	0.3	0.1	0.1	0.4	0.5	0.6
Rest	0.10	0.10	0.13	0.12	0.20	..	..	..	..	0.06	..	..	..	..	..	..	..	..

* Hebrides (Wager, 1956, p. 231); Hawaii (Wager, 1956, p. 226); Skaergaard (Wager and Deer, 1939, p. 218); Elephant's Head-New Amalfi (Poldervaart, 1944, p. 116)

on this type of diagram in Figure 20. The original and two subsequent liquids of the Skaergaard intrusion (Wager and Deer, 1939, p. 218; Wager, 1956, p. 220) and analyses of the Elephant Head-New Amalfi intrusion (Poldervaart, 1944, p. 116) are also shown for comparison. The three series are similar in showing marked iron enrichment. Averages of Ahr Lake at each stage form a smooth curve parallel to the Skaergaard trend but slightly displaced in the direction of higher iron ratio. The Elephant's Head-New Amalfi intrusion follows a similar course. Compared stage by stage (Table 13) the series show significant differences. Not all substages are represented by available analyses; hence allowances must be made for differentiation when comparing averages at different substages. The Skaergaard intrusion is richer in aluminum and sodium at each stage, and poorer in silica, than the Ahr Lake series. Potassium is similar in the early and middle stages but higher in the Skaergaard in the late stage. Iron is similar throughout but slightly more reduced in the middle and late stages of the Labrador series. The chemistry of the Elephant's Head-New Amalfi intrusion is similar to that of the Ahr Lake series at the two stages where they can be compared. Titanium and potassium are higher in the Elephant's Head-New Amalfi intrusion at both stages, silica is similar in the early stage but higher in the late-middle stage. Lime and magnesia are similar in both series in the early stage but decline more rapidly in the Elephant's Head-New Amalfi intrusion.

Data on Hebridean and Hawaiian tholeiites (Wager, 1956, p. 226, 231) are included in Table 13. The main features distinguishing Ahr Lake from Hebridean tholeiites are: (1) potassium is higher in the middle and late stages of the Hebridean series; (2) iron is similar in the early stage but increases rapidly in the Ahr Lake series in the final stage; it is more reduced in the Ahr Lake rocks at all stages; (3) silica increases more rapidly with differentiation in the Hebridean basalts and is lower in the early stage and higher in the late stage; (4) alumina decreases more rapidly in the Ahr Lake series and is higher in the early stage and lower in the late stage; (5) sodium varies little with differentiation in the Hebridean series and, except for the final stage, is higher than in the Ahr Lake series.

Points (4) and (5) reflect greater plagioclase fractionation in the Ahr Lake series. The more reduced state of Ahr Lake iron probably accounts for its extreme enrichment in the late stages. Lower potassium, silica, and sodium in the initial Ahr Lake magma distinguish it from initial Hebridean magma. Hawaiian tholeiites are similar to Hebridean tholeiites but have a more limited range of differentiation. They differ from the Ahr Lake series in the same way as the Hebridean tholeiites, except that silica is higher in Hawaiian rocks in both early and middle stages, and titanium is higher throughout the Hawaiian series.

Summary and Conclusions

The Labrador trough is composed of northwesterly trending sedimentary and volcanic rocks bounded on both sides by gneisses and schists. Sedimentary rocks predominate in the western half of the trough, volcanic rocks in the eastern half. Swarms of gabbro sills and a few ultramafic sills intrude the eastern portion of the trough throughout most of its length. Strata of the trough overlie gneisses unconformably on the western side and dip gently east. On the eastern side volcanic rocks are in fault contact with schists and gneisses. Folds within the trough are overturned to the southwest, and thrust faults dip northeast. The metamorphic grade increases from west to east across the trough and into the schists and gneisses beyond.

The present study is devoted to the mafic igneous rocks in the Ahr Lake area, New Quebec, which lies across the igneous belt on the northeast side of the trough.

The Ahr Lake area may be divided into two parts. Rocks of the western part are little metamorphosed and predominantly sedimentary; those of the eastern part are in the greenschist facies and predominantly volcanic. Gabbro sills intrude rocks of both parts.

Four varieties of gabbro sills are present: normal gabbro, metagabbro, leopard rock, and leuco- and melanometagabbro. Metagabbro is the metamorphosed equivalent of normal gabbro.

Many of the normal gabbro sills are differentiated. Olivine-bearing gabbros near the base of the sills grade up through even-grained gabbros, without olivine, to quartz- and micropegmatite-bearing gabbros near the top. One sill shows a differentiation from even-grained gabbro to albite gabbro pegmatite. The latter contains a finely crystalline mesostasis, interpreted as devetrified glass, but rare micropegmatite.

Salient features of the normal gabbro series are: The clinopyroxene trend initially follows

Hess's normal trend but rises to higher calcium contents in later pegmatitic gabbros. Plagioclase ranges from An_{80} to An_2; the normal range is An_{80-40}, with plagioclase from albite gabbro pegmatite at An_{10-2}. Iron-titanium oxides increase in amounts throughout the series, from 1 to about 10 per cent. An iron-rich chlorite with orange-brown to greenish pleochroism and high birefringence occurs in in the pegmatitic phases. The series is characterized by extreme iron enrichment and low potassium content. The differentiation trend is almost identical to the first part of the Skaergaard trend, but the Ahr Lake series shows no alkali enrichment. The ratio FeO/Fe_2O_3 for the series is unusually high and similar to that of the Skaergaard. Strontium content is unusually low; otherwise variation of trace elements resembles that in other tholeiitic series.

Major conclusions are: basaltic lava of the area approaches the composition of parent magma of the normal gabbro series, and differentiation occurred *in situ*. The magma was water-deficient and followed a differentiation trend of extreme iron enrichment. Alternative end products, gabbro pegmatite and albite gabbro pegmatite, probably resulted from slight differences in volatile contents. The more anhydrous magma produced albite gabbro pegmatite and completed its solidification with an interstitial glass phase. Origin of the albite is uncertain, but it is tentatively considered to be metamorphic.

Leopard rock is a coarse-grained felspathic gabbro spotted with spherical aggregates of plagioclase, 6–15 cm across. Sills of leopard rock have marginal zones, 30–75 feet wide, of common gabbro with sparsely scattered phenocrysts and aggregates of plagioclase. Chilled contact zones are dark basalt without phenocrysts. A number of leopard sills are composite, with normal gabbro occupying the centers. The normal gabbro interior may have differentiated in the same manner as other normal gabbro sills. Notable features of leopard rock are: Plagioclase is saussuritized and ranges from An_{50} to An_{70}, average about An_{60}. The composition of the chilled margin is similar to that of the parent gabbro magma disregarding introduced potassium. The strontium content of gabbro at borders of leopard-rock sills is similar to that of normal gabbros and suggests that they are related. The composition of leopard rock approximates that of a mixture of normal magma and the feldspathic aggregates of leopard rock.

Conclusions on the origin of leopard rock are: Leopard rock was formed by plagioclase enrichment in magma similar to the regional parent magma. Enrichment may have resulted from partial melting of the earth's mantle under conditions of very high pressure or of moderately high water pressure. Crystallization before final intrusion yielded a magma rich in plagioclase crystals. Continued partial melting may have produced normal basaltic magma. This mode of origin agrees with the observed sequence of intrusion: leopard rock–normal gabbro. Local concentration of plagioclase by currents in a magma reservoir is also possible and would explain why leopard rock is a mixture of plagioclase and parent magma and why leopard rock was intruded before normal gabbro. The extent and quantity of leopard rock in the Labrador trough favors the first mode of origin. The magma with its load of suspended plagioclase assumed a flow structure upon intrusion, so that plagioclase was concentrated in a central sheet, and a thin layer at each margin was left nearly free of plagioclase. This is called sheet flow, and it accounts for the distribution of feldspathic aggregates within leopard-rock sills. Composite sills were formed when normal gabbro intruded the semiconsolidated centers of leopard rock.

Leucometagabbro and melanometagabbro are unusual varieties of metagabbro which have no equivalents among the normal gabbros. They are found together in two sills; melanometagabbro occupies the lower portion and leucometagabbro the upper. In one of the sills, and possibly in both, they are gradational across a narrow zone. Melanometagabbro consists of coarse-grained amphibole ophitically enclosing altered plagioclase. Relic textures indicate that it was originally composed of pyroxene and olivine, with subordinate plagioclase. Leucometagabbro consists of albite and epidote with minor amphibole. The two rocks have similar mafic indices and are equally sodic.

The origin of leuco- and melanometagabbro is uncertain. Division of sills into a lower portion consisting of the heavier constituents, and an upper portion consisting of the lighter constituents, of a mafic magma suggests that gravity was a factor in their origin, but the mechanism is obscure. Possibly in a crystal mush during intrusion, heavier crystals work their way to the bottom and displace lighter crystals upward, so that eventually separation of light and heavy constituents would be nearly complete.

The normal gabbros are typical tholeiites. The parent magma is similar in composition to the average of a number of tholeiitic magmas (Nockolds and Allen, 1956, p. 76), but the series has distinctive features. Potassium and strontium are lower than in most tholeiitic provinces, and differentiation leads to extreme iron enrichment. Compared with the Skaergaard and Elephant's Head-New Amalfi intrusions, with similar differentiation trends, the Ahr Lake series shows consistently lower aluminum and sodium and higher silicon than the Skaergaard intrusion and lower titanium and potassium than the Elephant's Head-New Amalfi intrusion. Differences are also noted between the Ahr Lake series and Hebridean and Hawaiian tholeiites.

Regional metamorphism decreases westward across the Ahr Lake area from the quartz-albite-epidote-biotite subfacies of the greenschist facies in the northeast corner, through the quartz-albite-muscovite-chlorite subfacies between Keato and Murdock lakes, to subgreenschist-facies rocks west of Murdock Lake. Local occurrences of pumpellyite and prehnite in the western zone are interpreted as indicating zeolite facies. Albite of albite gabbro pegmatite may be due to development of zeolite-facies assemblage in rocks with a glass phase.

Effects of contact metamorphism are slight. Sedimentary rocks adjacent to gabbro sills are recrystallized to finer grain size, and some show meager development of biotite.

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EVOLUTION OF TILL-STONE SHAPES, CENTRAL NEW YORK

BY CHAUNCEY D. HOLMES

ABSTRACT

A total of 3234 unselected till pebbles and cobbles from 27 localities in central New York are of three lithologic types: sandstones, limestones, and shale (including siltstones). The most distant localities are about 20 miles south of the chief limestone outcrops at the Allegheny Plateau escarpment. Changes in size, shape, and roundness with increasing distance of glacial transport show a relative decrease of about 25 per cent in wedge-form and in rhombohedroid stones and a relative increase of more than 100 per cent in ovoids. Other forms show little variation. Although wedge forms tend to be more conspicuously striated than others, the inference is that rounding predominated over faceting, and that prolonged transportation by the middle-latitude ice sheets tended to produce stones of ovoid form rather than wedge shapes as has been claimed. All three lithologic types show these tendencies, but in different degrees depending on relative toughness. Reduction of the larger sizes provided a continuing supply of smaller ones, and gradual depletion of the larger sizes is reflected in the changing cobble-pebble ratio downcurrent from the source outcrops. Here also, rates of change are governed largely by relative resistance. However, with increasing distance of transport, cobbles increase in roundness more than pebbles, indicating less susceptibility to crushing and perhaps a faster rate of rounding. Only about 10 per cent of the till stones attained a degree of roundness greater than 0.4 on Krumbein's scale (ρ on Holmes's scale).

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Introduction

Belief that glacially transported stones tend to develop a characteristic or diagnostic shape is supported to some extent by the few published reports on this subject; but most studies have not included the full context from which conclusions should be drawn. Several European geologists of the last century contributed many incidental and important qualitative observations on till-stone features. Perhaps the most significant contribution to this problem was made by von Engeln (1930). He concluded that glacial abrasion produces primarily a faceted, flatiron-shaped end product, developed probably from original "rectangular prisms or cubical blocks" of rock. He referred also to a secondary type which shows rounding by abrasion but lacks facets.

Using somewhat different quantitative data, Wentworth (1936) concluded that the characteristic shape produced by glacial abrasion is that of a "pentagonal wedge" as viewed in profile from the top. This appears to be essentially von Engeln's flatiron shape. Both von Engeln and Wentworth emphasized prominent striation as indicating extensive shaping by the glacier. The fact that wedge-shaped stones generally carry more conspicuous striae than those of other shapes apparently supports their thesis. From till-fabric analysis, the present writer (1941, p. 1333) has suggested that the better development of striae on wedge-form stones is chiefly a consequence of their initial triangular form, their striae engraved upon pre-existing surfaces. The question of shaping by glacial abrasion has needed further study.

The present report is based on a study of 3234 pebbles and cobbles taken from unleached till at 27 localities, mostly in the Cazenovia and Tully (15-minute) quadrangles, in central New York. The data were collected in 1937–1938 incidental to the till-fabric study (Holmes, 1941). The New York area is exceptionally favorable for this type of investigation because the glacier flowed generally southward across the strike of the strata which contributed to the glacier load. Distances of transportation can therefore be closely approximated, and progressive changes in till-stone characteristics, statistically considered, can be evaluated.

Regional Setting

The Cazenovia and Tully quadrangles lie within the Allegheny Plateau (Fig. 1), their northern borders extending almost to the plateau escarpment which rises above the Ontario Lowland to an average height of about 500 feet. Middle Devonian limestones (Onondaga and "Helderberg"), underlain by Upper Silurian limestones and dolomites, form the major element of the escarpment, and their outcrops extend into the northern portions of the two quadrangles. Various units within the overlying Hamilton group (mostly shales and siltstones) form minor discontinuous or local escarpments across the central part of the area. Across the southern portion the Portage escarpment (siltstones and fine-grained sandstones) rises abruptly along an irregular zone with several hilltop outliers. All strata dip gently to the south or southwest, which partly obscures the scarped aspect of the plateau topography. Local relief varies much but averages about 500 feet. The highest hilltops rise slightly above 2000 feet, and the maximum relief is about 1600 feet.

A few deep valleys lead northward onto the Ontario Lowland from the major drainage divide which crosses the southern portion of the area. Along these valleys the glacier carried drift from the lowland much farther south than it did along the intervening uplands (Holmes, 1952). Thickness of drift ranges from a discontinuous veneer on some hill slopes to a maximum of about 1000 feet in the deepest valley.

Three distinctive lithologic groups make up the bulk of drift in the plateau: quartzitic sandstones, carbonate rocks, and a varied shale and siltstone suite. Nearly all the carbonate rocks came from the limestones and dolomites at the escarpment. Small amounts represent the thin Silurian calcareous beds cropping out on the Ontario Lowland, and a still smaller quantity, of Ordovician age, was brought out of the Lake Ontario basin. Within the plateau the Tully limestone supplied minor quantities. In this study the term limestone is used for the entire carbonate group.

Most of the quartzitic sandstones are of Oswego and Medina ages and crop out along the south shore of Lake Ontario. Their distinctive red and green colors and medium-grained texture are the chief criteria for identification. With them in the plateau drift are a variety (small in total quantity) of finer-grained sedimentary quartzites of pre-Ordovician age and a few coarse-grained fragments from the Oriskany formation (or basal Onondaga) which crops out along the escarpment.

All the identifiable shales and siltstones in the plateau drift were derived locally from the Hamilton, Genesee, and Portage strata. They are by far the softest and weakest of the three lithologic groups. In this paper the term shale denotes also the siltstones.

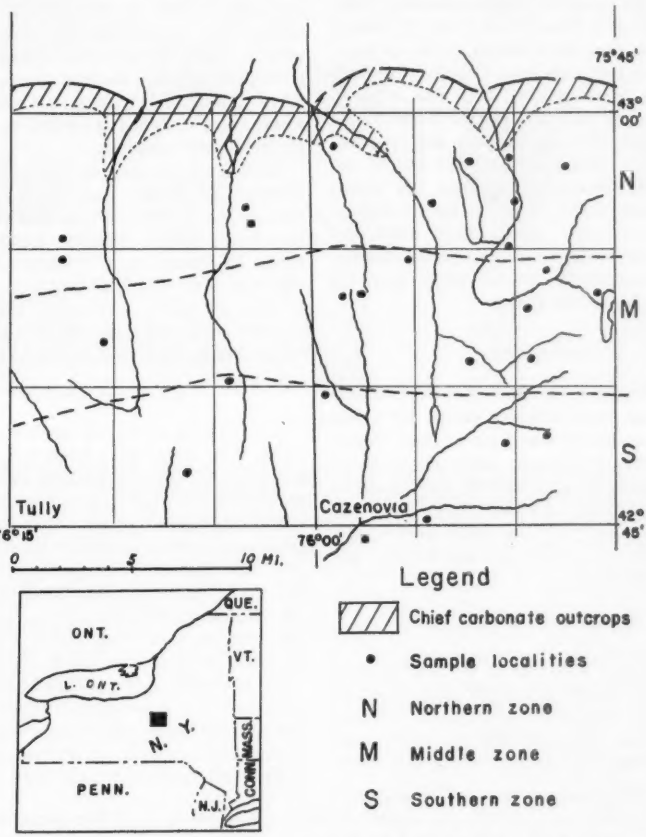

FIGURE 1.—SKETCH MAP OF CAZENOVIA AND TULLY QUADRANGLES IN RELATION TO LIMESTONE-DOLOMITE OUTCROP ZONE WHICH FORMS THE NORTHERN BORDER OF THE ALLEGHENY PLATEAU IN THE REGION

ORGANIZATION OF THE PRESENT STUDY

Sample-Locality Groups

As shown in Figure 1, the 27 localities from which samples were taken have been divided into three groups representing increasing distance southward in the plateau. Group N (northernmost) includes 11 localities located not more than 4 miles from the limestone outcrops along the escarpment. The middle group (M) contains 9 localities ranging from 5 to 10 miles from the outcrops; and the southernmost group (S), with 7 localities, ranges from 10 to 20 miles south of the escarpment. Till stones in the three regions show strong similarities; the analyses which follow reveal some consistent differences.

Of the three lithologic groups, the sandstones were carried farthest; their outcrops are about 35 miles north from this segment of the escarpment. Except for these sandstones and a small per cent of the carbonates, the till stones studied were transported not more than 20 miles.

Till-Stone Forms or Shapes

Each of the 3234 stones was classified as to form in one of six categories: ovoid, discoid, tabular, wedge-form, rhombohedroid, and varihedroid. (*See* Figure 1 of Plate 1.) Except for the last two, these names are in common use. Rhombohedroids have (ideally) two sets of parallel or subparallel sides like a brick, although much variation occurs. Varihedroids have surfaces with such variance and irregularity that they cannot be assigned any of the other forms. Subsequent analysis has shown that tabular and discoid stones constitute essentially a single group, their differences being chiefly in degree of roundness (Holmes, 1941). Consequently in this study they are combined as discoid-tabular.

Sizes of Stones

Stones having a minimum dimension of less than 0.5 cm were rejected, except for a few stones whose length was 3 cm or more. Maximum dimensions (length) range up to about 25 cm. Stones up to 6 cm long are classed as pebbles, and those 7 or more cm in length are listed as cobbles. Of the total 3234 stones analyzed in this study, 202 are cobbles and 3032 are pebbles.

Roundness of Stones

Four degrees of roundness are used: *a* (well rounded); *b* (moderately rounded); *c* (slightly rounded); *d* (sharply angular). These compare approximately with published scales, although no field measurements of roundness were made.

This study	*Wentworth* (1923)	*Tester* (1931)	*Krumbein* (1941)
d	0.00–0.01	0–18	0.1–0.2
c	0.02–0.20	19–45	0.3–0.4
b	0.21–0.30	46–75	0.5–0.6
a	0.31–1.00	76–100	0.7–0.9

Roundness *d* and *c* are by far the most common (Tables 3, 5); till stones of roundness *a* are rare. However, many stones have one or more surfaces inherited from a larger one, generally discoid or ovoid, which had been broken after having attained roundess *b* or *a*. Others, broken while still of *c* roundness, show correspondingly less contrast in edge characteristics. Unless the breakage was slight, broken stones were classified according to the degree of subsequent rounding.

Although form and roundness are distinctly separate properties, they overlap to some extent. The term ovoid implies a fair degree of roundness, and a sharply angular "ovoid" would be classed as a varihedroid. Similarly, rounding of a varihedroid past the *b* stage brings it near, or within, the ovoid group. Consequently only three degrees of roundness are recognized in each of these two form groups.

Lithologic Selection

For purposes of the till-fabric study in 1938, every stone larger than minimum size at each locality was taken regardless of other characteristics until the required number (100 or more) had been collected. This total included a small number of gneisses and allied types and a few chert fragments. Altogether these made up less than 2 per cent of the total. They are too few in number to treat as separate groups in the present study, and their characteristics naturally exclude them from the three dominant lithologic suites. Therefore they have been omitted from consideration. All other stones from the several localities have been included, and the entire assemblage for analysis is therefore essentially unselected.

Analysis and Interpretation

Introduction

Texture of the till varies somewhat over the area, but the histogram (Fig. 2) is fairly

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PLATE 1.—FORMS AND ROUNDNESS IN TILL STONES

FIGURE 1.—Typical till stones illustrating forms and degrees of roundness. About one-fourth natural size. All have traces of original concave surfaces. Note low value of *c* roundness and close approach of numbers 4 and 8 to ovoid form. Roundness, *d* to *a*. Discoid-tabular: 1 *c*, 2 *c*, 3 *d*, 12 *b*; rhombohedroids: 4 *b*, 5 *c*, 6 *c*, 7 *d*; Wedge-form: 8 *b*, 9 *c*, 10 *c*, 11 *d*; ovoids: 13 *a*, 14 *b*; varihedroids: 15 *c*, 16 *c*, 17 *d*.

FIGURE 2.—Well-rounded (*a*) boulder of "Cobleskill" dolomite on east shore of Skaneateles Lake, about 6 miles west of Tully Quadrangle. Size, about 6 by 4 by 4 feet. It lies approximately 10 miles from its probable source ledge and 250 feet higher in elevation. Its present ovoid form was evidently developed from an original rhombohedroid by glacial abrasion and spalling in transit. Its surface is intensely pressure-scarred and deeply striated.

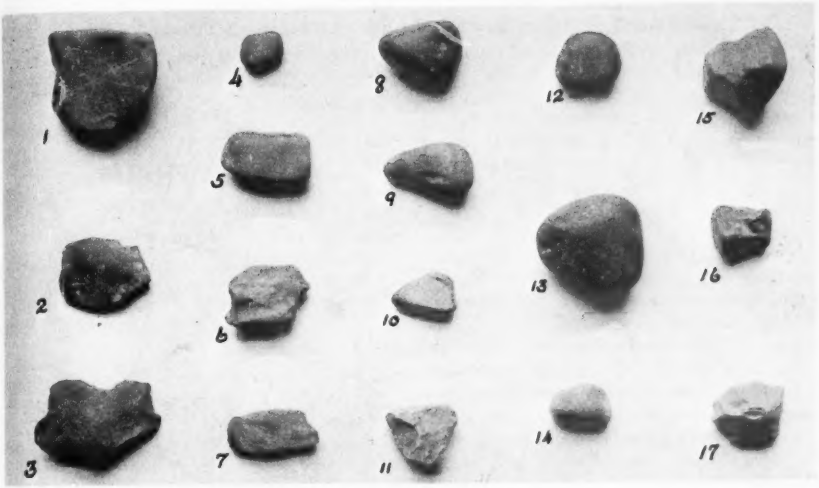

FIGURE 1

FIGURE 2

FORM AND ROUNDNESS IN TILL STONES

representative of the localities sampled. The stones collected represent approximately the coarser two-thirds of the pebbles and almost all the cobbles. By weight the pebbles and

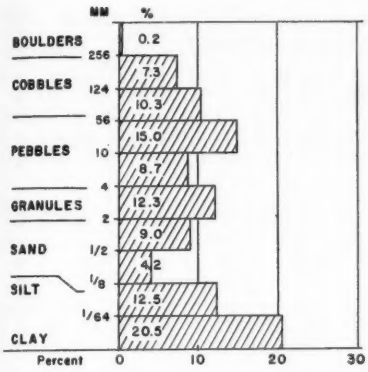

FIGURE 2.—COMPOSITE HISTOGRAM (THREE LOCALITIES) OF TILL IN CAZENOVIA AND TULLY QUADRANGLES

cobbles are nearly equal, but in numbers the cobbles make up less than 7 per cent of the total. Analysis is based on numerical values.

Till stones at any given exposure have had varied histories. Some have undergone change only by abrasion in being carried from their source ledges, which may have been only a few feet or many miles distant. Most are apparently fragments produced by crushing of larger stones at some intermediate distance between source and depositional site, with subsequent abrasion through the remainder of the distance. Therefore any statistically observed trends in shape and roundness with increasing distance of transport represent the cumulative dominance of subsequent changes over initial angularity from crushing. Crushing of stones in transit—the making of new sharply angular shapes—tends to mask the effects which glacial abrasion has in altering the shapes of stones. Statistical comparisons therefore give minimum values of changes due to abrasion.

Plan of Analysis

Table 1 lists all the data on the total number of till stones, classified as to form, roundness, size, and lithology, with increasing distance southward in the plateau. Data selected from Table 1 are then presented in simpler tables for easier and more direct comparisons of significant trends in till-stone characteristics, which are discussed in connection with the appropriate tables.

The problem of whether there is a "type form" toward which glacial abrasion tends is best approached by considering certain background data on size and lithology in relation to shape and roundness. Significant comparisons are given in Tables 2–4.

Lithology of the Drift

Table 2 shows the changes in relative per cent of the three lithologic suites southward in the plateau. The most noticeable change is an increase in per cent of shale, as would be expected because these rocks make up most of the plateau strata. With this increase goes a marked decrease in the per cent of limestones. These changes suggest a dilution of the glacier load by addition from the plateau rocks. This means that the total load would have been greatly increased as the glacier moved southward, which may have occurred to some extent. However, this dilution hypothesis is opposed by the fact that the relative per cent of sandstones is increased in significant amount, instead of decreased, through the same distance.

The sandstones are far tougher than the other rocks and therefore survive the processes of crushing and abrasion better than the others. Moreover, very few sandstone boulders occur in the drift even at the plateau escarpment, which could have furnished a continuing supply of pebble-cobble sizes from crushing as transportation continued. In contrast, many limestone boulders, large and small, occur through the first few miles of southward transport from their outcrops. Crushing and abrasion of these larger sizes must have replenished the smaller to a considerable extent. Had limestone boulders been few or absent, the southward decline in per cent of carbonates would have been much greater.

Throughout the plateau the till is reasonably uniform in grade-size distribution. This suggests that the finely ground products of crushing and abrasion were not retained in the glacier load which was finally concentrated in the end moraines. These considerations support the interpretation of gradual or progressive reduction of larger sizes and disappearance of the smaller sizes from the glacier load, at a rate inversely proportional to the toughness of the stones. This interpretation is further supported by observing the field occurrence of the distinctive black Genesee shale fragments in the till. This shale is notably soft, and till near the outcrop contains large quantities of

TABLE 1.—COMPOSITE DATA ON 3234 TILL STONES IN CENTRAL NEW YORK

Till stones are classified as to form, roundness, size, and lithology, with increasing distance southward in the plateau. Numbers in parentheses indicate striated stones in the particular group: for example, 100 (34) means 100 stones, 34 of which are striated. Locality groups or zones: N (northern), M (middle), S (southern)

Form	Roundness	Size	N			M			S		
			Limestone	Shale	Sandstone	Limestone	Shale	Sandstone	Limestone	Shale	Sandstone
Rhombohedroid	*d*	Peb.	100 (34)	14 (0)	9 (0)	19 (3)	42 (6)	6 (0)	8 (4)	28 (0)	4 (0)
		Cob.	8 (3)	1 (0)	1 (0)	2 (2)	6 (3)	0	1 (1)	0	0
	c	Peb.	83 (41)	24 (5)	19 (2)	92 (45)	102 (13)	21 (0)	19 (12)	48 (6)	17 (0)
		Cob.	12 (6)	0	2 (0)	9 (6)	5 (1)	2 (0)	0	3 (0)	1 (0)
	b	Peb.	4 (3)	3 (0)	2 (0)	9 (7)	6 (1)	5 (0)	1 (1)	0	3 (0)
		Cob.	1 (0)	0	1 (0)	0	1 (0)	0	0	0	0
	a	Peb.	0	0	0	0	0	0	0	0	0
		Cob.	0	0	0	0	0	0	0	0	0
Wedge-form	*d*	Peb.	207 (72)	33 (0)	24 (0)	80 (17)	105 (13)	11 (0)	21 (6)	62 (0)	11 (0)
		Cob.	12 (6)	1 (0)	2	2 (1)	5 (2)	0	1 (0)	3 (1)	1 (0)
	c	Peb.	180 (97)	26 (1)	22 (0)	154 (72)	141 (24)	33 (0)	56 (33)	76 (7)	21 (0)
		Cob.	19 (15)	1 (0)	0	12 (6)	5 (2)	4 (0)	5 (2)	2 (2)	1 (0)
	b	Peb.	2 (0)	0	2 (0)	13 (7)	5 (1)	6 (0)	1 (1)	2 (0)	3 (0)
		Cob.	0	0	0	2 (1)	2 (0)	0	0	0	0
	a	Peb.	0	0	0	0	0	0	0	0	0
		Cob.	0	0	0	0	0	0	0	0	0
Discoid-Tabular	*d*	Peb.	55 (16)	9 (1)	15 (0)	42 (14)	74 (7)	11 (0)	30 (19)	24 (3)	10 (0)
		Cob.	3 (1)	1 (0)	0	1 (1)	4 (1)	0	0	0	2 (0)
	c	Peb.	20 (10)	9 (5)	5 (1)	37 (23)	39 (4)	12 (0)	9 (6)	23 (5)	4 (0)
		Cob.	1 (1)	1 (1)	0	3 (3)	2 (0)	1 (0)	1 (1)	2 (2)	1 (0)
	b	Peb.	3 (1)	5 (1)	4 (0)	15 (8)	5 (2)	5 (0)	4 (3)	4 (2)	1 (0)
		Cob.	0	0	0	1 (0)	1 (0)	0	0	0	1 (0)
	a	Peb.	0	0	1 (0)	0	0	1 (0)	0	0	2 (0)
		Cob.	0	0	0	0	0	0	0	0	0
Varihedroid	*d*	Peb.	69 (29)	10 (1)	5 (0)	24 (4)	34 (2)	1 (0)	8 (3)	29 (0)	6 (0)
		Cob.	6 (2)	0	0	0	0	0	0	1 (0)	0
	c	Peb.	48 (24)	3 (0)	8 (0)	17 (10)	31 (3)	10 (1)	19 (10)	21 (2)	5 (0)
		Cob.	5 (4)	0	0	2 (1)	4 (1)	0	1 (0)	2 (0)	0
	b	Peb.	0	0	0	0	0	1 (0)	0	0	1 (0)
		Cob.	0	0	0	1 (1)	0	0	0	0	0
Ovoid	*c*	Peb.	2 (1)	2 (0)	3 (0)	15 (7)	10 (2)	7 (0)	6 (5)	7 (3)	4 (0)
		Cob.	0	0	1 (0)	1 (1)	0	1 (0)	1 (1)	0	1 (0)
	b	Peb.	19 (8)	0	19 (0)	32 (15)	11 (1)	22 (0)	21 (12)	14 (2)	14 (0)
		Cob.	1 (0)	1 (1)	1 (0)	2 (1)	0	2 (1)	1 (1)	3 (2)	2 (0)
	a	Peb.	0	0	3 (0)	0	2 (0)	5 (0)	0	0	1 (0)
		Cob.	0	0	0	0	0	0	0	0	0

it in pebble size or larger; but none of it survived in recognizable fragments farther than about 3 miles from the source ledges.

Evidence that the finer constituents of the glacier load disappeared while the coarser were kept in transport is significant because it affords an interpretation of the transportation environment. Beginning with the reasonable assumption that the glacier which deposited the plateau drift was of the temperate category—that is, the temperature in the basal zone was at the pressure-melting point—heat from friction of ice movement and especially from abrasion of the glacier load would have produced much meltwater in addition to that which existed momentarily from pressure melting. This additional water would have moved downward, and with it (though perhaps less rapidly) the finer products of abrasion and crushing would have been brought to the base of the glacier and deposited as lodgement drift. Larger fragments at the interface that were caught in the deposited material would also escape from the ice. Larger particles above the interface would be more nearly encased in the ice and thus kept from settling with the intergranular meltwater.

The tentative conclusion to be drawn thus far is that glacial erosion of bedrock blocks, large and small, occurred as the glacier moved southward. These blocks were abraded and crushed, and the smaller particles of all types were selectively lost from the glacier load about as fast as new load was acquired. The toughest stones survived best and were carried farthest.

TABLE 2.—THE THREE LITHOLOGIC SUITES, UNDIVIDED, WITH INCREASING DISTANCE SOUTHWARD IN THE PLATEAU

In per cent. Number of stones in each group shown in parentheses. Locality groups or zones: N (northern), M (middle), S (southern)

	N	M	S
Limestone	(860) 74.6	(587) 42.0	(214) 31.2
Shale	(144) 12.4	(642) 46.0	(354) 51.7
Sandstone	(149) 13.0	(167) 12.0	(117) 17.1
Total	(1153) 100.0	(1396) 100.0	(685) 100.0

Till-Stone Sizes

Changes in cobble-pebble ratio with increasing distance of transport shed additional light on the comparative response of the three lithologic types to the transportation environment. For the entire area, cobbles make up about 6 per cent of the total (Table 3). Percentages in the shale-siltstone group show relatively fewer cobbles than average and little change with increasing distance (Table 3). This reflects the continuing supply of material eroded from the local bedrock and probably also a relatively faster rate of reduction from cobble size, presumably by crushing. Limestone cobbles, however, show an appreciable decline (7.9–5.1) per cent) which may be safely attributed to depletion of limestone boulders in the glacier load with increasing distance from their outcrops.

As in Table 2, the sandstones show a trend opposite that of the limestones, with an increase in relative abundance of cobbles, from 5.4 to 8.6 per cent. Sandstone boulders are extremely scarce in the plateau drift and therefore could not have supplied many additional fragments of cobble size. Again, the superior resistance of these rocks offers a reasonable interpretation. Field observation of broken sandstones (from crushing) showed that the larger stones generally broke into two pieces instead of many, as is common with both limestones and shales. A large cobble could thus make two smaller cobbles instead of a number of pebble-sized fragments. Moreover, with destruction of the larger limestone boulders, the surviving sandstone cobbles were evidently more resistant than anything else in the glacier load (except perhaps some of the few granitic rocks) and were therefore progressively less subject to crushing.

Another probable factor in the increased ratio of sandstone cobbles to pebbles is that the sandstone pebbles, being smaller, were relatively more vulnerable to crushing.

Thus the tentative conclusions drawn from changes in lithologic percentages are supported also by the changing pebble-cobble ratios.

Roundness

Whereas shape of a till stone produced by crushing can be interpreted with fair assurance

by reference to the configuration of partial surfaces inherited from the larger stone of which it was a part, roundness is more obscure in history of development. Rounding by glacier transport has long been recognized, but the possibility must not be ignored that a given stone may have had an episode of rounding by water transport before incorporation into the glacier load. Criteria for distinguishing between glacier-rounded stones and water-rounded stones are necessarily tenuous but nevertheless suggestive. Almost all well-rounded till stones retain traces of original concave surfaces to a far greater extent than is generally observed on beach or stream pebbles and cobbles. The degrees of till-stone roundness show gradual transition from the sharply angular condition, suggesting that the same process by which roundness *c* developed from *d* is continued in further rounding to the *b* and *a* stages. Comparative percentages are given in Table 3. Undoubtedly some stones in the till have been rounded by earlier water action, but their number and the amount of rounding are believed not to invalidate qualitatively the inferences to be drawn from the comparative percentages.

TABLE 3.—TILL-STONE SIZES CLASSIFIED AS TO LITHOLOGY AND ROUNDNESS, WITH INCREASING DISTANCE SOUTHWARD IN THE PLATEAU

In per cent. Number of stones in each group shown in parentheses. Locality groups or zones: N (northern), M (middle), S (southern)

Size	(1661) Limestone			(1140) Shale			(433) Sandstone			Total
	N	M	S	N	M	S	N	M	S	
Pebbles	(792) 92.1	(549) 93.5	(203) 94.9	(138) 95.8	(607) 94.6	(338) 95.5	(141) 94.5	(157) 94.0	(107) 91.4	(3032) 93.8
Cobbles	(68) 7.9	(38) 6.5	(11) 5.1	(6) 4.2	(35) 5.4	(16) 4.5	(8) 5.4	(10) 6.0	(10) 8.6	(202) 6.2
Total	(860)	(587)	(214)	(144)	(642)	(354)	(149)	(167)	(117)	(3234)

Roundness	Pebbles			Cobbles			Average	
	N	M	S	N	M	S	Pebbles	Cobbles
d	(550) 94.0	(449) 95.7	(241) 96.4	(35) 6.0	(20) 4.3	(9) 3.6	(1240) 95.1	(64) 4.9
c	(454) 91.5	(721) 93.4	(335) 94.1	(42) 8.5	(51) 6.6	(21) 5.9	(1510) 93.0	(114) 7.0
b	(63) 92.6	(135) 91.9	(69) 90.8	(5) 7.4	(12) 8.1	(7) 9.2	(267) 92.0	(24) 8.0
a	(4) 100.0	(8) 100.0	(3) 100.0	(0) ..	(0) ..	(0) ..	(15) 100.0	(0) ..
Total	(1071) 92.9	(1313) 94.1	(648) 94.6	(82) 7.1	(83) 5.9	(37) 5.4	(3032) 93.8	(202) 6.2

About half of the total number of stones are in the *c* groups, and nearly as many are in *d*. In further rounding from *c* to *b* the numbers are sharply reduced, and an even more drastic decline occurs between *b* and *a*. Less than 0.5 per cent of the total number of stones are of *a* roundness (Table 5). These percentages probably indicate that the rounding process is

most likely to be terminated by crushing before it goes much past the *c* stage. However, many limestones of *a* roundness still retain traces of original concave surfaces, highly polished and with exceedingly fine striae. These same features are more common on similar stones of *b* roundness. The relationships are believed to be good evidence that glacial transport can and does cause rounding to a high degree if the stone escapes crushing, and that the observed degrees of roundness are chiefly the effects of glacial abrasion.

Further data on degree of roundness attained before fracturing are given in Table 4 and in Figure 3. Fragments of discoids and ovoids which had reached roundness *a* or *b* can be recognized readily and were recorded in the field notes. Subsequent rounding of these fragments was slight, and most were classified as *d* roundness. None had acquired subsequent rounding past the *c* stage. They occurred singly and widely spaced in the drift, and at no outcrop could any two of them be fitted together. Moreover, the collecting sites were widely spaced. Therefore all are presumed to represent different individual rounded stones, most of which are limestones and sandstones. Whereas only 15 ovoid-discoid pebbles of roundness *a* (Table 1) were collected intact, 53 (Table 4) had been broken shortly before they were deposited. Similar comparisons within the *b* groups are significant although less striking (198 intact, 148 broken).

TABLE 4.—ROUNDED PEBBLES BROKEN IN TRANSIT

Only discoids and ovoids of roundess *b* and *a* are recorded because others cannot be recognized with assurance. Numbers of stones in each group shown in parentheses. Locality groups or zones: N (northern), M (middle), S (southern)

		N	M	S
Limestones				
Discoid	*b*	(2)	(8)	(8)
	a	(0)	(1)	(1)
Ovoid	*b*	(14)	(27)	(16)
	a	(4)	(7)	(5)
Shale				
Discoid	*b*	(0)	(6)	(9)
	a	(0)	(1)	(2)
Ovoid	*b*	(3)	(7)	(12)
	a	(0)	(7)	(2)
Sandstones				
Discoid	*b*	(4)	(5)	(2)
	a	(1)	(1)	(5)
Ovoid	*b*	(6)	(6)	(13)
	a	(5)	(4)	(7)
Total	*b*	(29) 71.8	(59) 73.7	(60) 73.2
(per cent)	*a*	(10) 28.2	(21) 26.3	(22) 26.8
Total		(39)	(80)	(82)
Per cent in pebble group		3.6	6.1	12.7

Table 4 shows, ratio between *a* and *b* roundness in these stones before they were broken is roughly 1 to 3 and remains remarkably constant southward in the plateau. For the unbroken discoid-ovoid pebbles of comparable roundness (Table 1), those of roundness *a* are considerably less common. In the northern group the ratio of *a* to *b* is 1:13; middle group, 1:12; southern group, 1:19. These ratios suggest that stones of roundness *a* are much more susceptible to fracturing than are those of roundness *b*. Till-fabric studies (Holmes, 1941) have led to the conclusion that rounding is progressively conducive to rolling during glacial transportation, which in turn probably causes faster rounding. Apparently the increased tendency to roll forward also makes the stone more likely to be crushed. The vertical space available for rotation may be limited by nonrotating larger stones both above and beneath, and presumably the superjacent stones move faster than the lower ones. If the space between is not great enough to accommodate the largest diameter of the rolling stone normal to its axis of rotation, crushing would seem to be inevitable. However, if a high degree of sphericity were attained, the likelihood of crushing would presumably be greatly reduced. (*See* Figure 3.)

Trends in developing rounded shapes are revealed also by comparing percentages of pebbles and cobbles. On the average, the cobble ratio increases with roundness, from 4.9 per cent (*d*) to 8 per cent in roundness *b*. In groups *d* and *c* the per cent of cobbles decreases southward, but in the *b* group it increases. This is perhaps another indication that the larger stones are less likely to be

crushed, and therefore the rounding process continues through a correspondingly greater distance of transport. (*See* Figure 2 of Plate 1.)

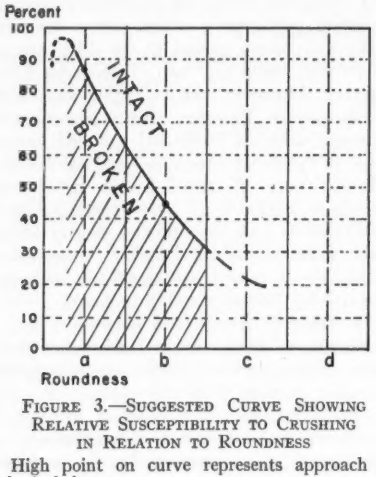

FIGURE 3.—SUGGESTED CURVE SHOWING RELATIVE SUSCEPTIBILITY TO CRUSHING IN RELATION TO ROUNDNESS

High point on curve represents approach to spherical shape.

Form and Roundness

Table 5 presents critical data on relative abundance of the several till-stone forms with increasing distance of transport, and the relationships are best seen by noting first the totals in which the three lithologic suites are combined. From north to south, rhombohedroids decrease about one-fifth, and wedge forms decrease about one-sixth. The most notable increase is in the ovoids—more than 150 per cent. Discoid-tabular forms also show notable increase. In roundness, *d* decreases significantly, and *b* increases nearly 100 per cent. Increase in *c* roundness is moderate.

The contrasts among the lithologic groups reflect differences in strength and hardness between the sandstones and limestones and the effect of continuing replenishment of the shale rocks. Percentages in form of the sandstones remain nearly static, but variations in roundness suggest that more crushing occurred in the southern zone than in the middle zone. Limestones contribute most to the totals; limestone ovoids increase more than 400 per cent. Rhombohedroids show the greatest decrease, with moderate decrease also in the wedge forms. Both limestone and sandstone varihedroids are least abundant in the middle zone and increase into the southern. A probable reason for this is that end moraines are more numerous in the southern zone, and the thinner and more rigid ice at the glacier terminus, with increasing concentration of the glacier load near the limit of transport, may have caused more crushing of the stones than that which occurred a few miles back from the glacier margin. Roundness percentages in the limestones and sandstones seem to support this interpretation. If this is the case, the actual rate of progressive changes in form and roundness with increasing distance of transport is more accurately shown by comparisons between the northern and middle zones than between the northern and southern.

In summary, the data in Table 5 show that with increasing distance of glacier transport, wedge-form stones decrease in relative abundance, although not so rapidly as rhombohedroids. Varihedroids decrease at a possibly still faster rate. The most notable increase is in the relative abundance of ovoids; increase is conspicuous also in the discoid-tabular group. A general increase in roundness occurs, especially in the limestones. Additional ovoids and discoids can be made only by rounding, while varihedroids and all other sharply angular forms observed in the field can be produced only by crushing. The conclusion indicated by these considerations is that glacial abrasion tends to convert rhombohedroids, wedge forms, and varihedroids to ovoid or discoid forms unless crushing interrupts the abrasion process.

Table 5 has shown the similarity of trends in relative abundance of wedge forms and rhombohedroids, and further comparison is desirable inasmuch as wedge forms have been thought to have evolved from (rhombohedroid ?) blocks (von Engeln, 1930). Accordingly, Table 6 shows variations in roundness for these two forms in the three lithologic suites. Numerically (Table 5), wedge forms are nearly twice as abundant as rhombohedroids. More than half of those in the northern zone are sharply angular shapes freshly crushed. This is especially true of the limestones, the dominant lithology in the northern zone. With southward transport, rhombohedroids show a slightly faster rate of rounding; a definitely larger per cent reached the *b* stage than is the case with the wedge forms. Trends in each category of roundness are essentially identical although variations occur among the lithologic groups. Especially, the relatively higher per cent of sandstones of *b* roundness is further evidence that stones of weaker lithologic types are lost from this roundness category largely by crushing.

TABLE 5.—THE THREE LITHOLOGIC SUITES CLASSIFIED AS TO FORM AND ROUNDNESS, WITH INCREASING DISTANCE SOUTHWARD IN THE PLATEAU

In per cent. Number of stones in column or group shown in parentheses. Total in each form group shown at right. Locality groups or zones: N (northern), M (middle), S (southern)

	(1661) Limestone			(1140) Shale			(433) Sandstone			(3234) Total			No.
	(860) N	(587) M	(214) S	(144) N	(642) M	(354) S	(149) N	(167) M	(117) S	(1153) N	(1396) M	(685) S	
Form													
Rhombohedroid	24.2	22.3	13.5	29.2	25.2	22.4	22.8	20.4	21.4	24.6	23.4	19.5	744
Wedge-form	48.9	44.8	39.3	42.3	40.8	41.0	33.6	32.3	31.6	46.1	41.5	38.8	1377
Discoid-tabular	9.5	16.9	20.5	17.4	19.4	15.0	16.8	18.0	17.9	11.4	18.2	17.2	504
Varihedroid	14.9	7.5	13.1	9.0	10.8	15.0	8.7	7.2	10.3	13.4	9.0	13.6	372
Ovoid	2.6	8.5	13.5	2.1	3.6	6.8	18.1	22.2	18.8	4.5	7.9	11.0	237
	100.1	100.0	99.9	100.0	99.8	100.2	100.0	100.1	100.0	100.0	100.0	100.1	3234
Roundness													
d	53.4	28.9	32.2	48.0	42.0	41.6	37.6	17.4	29.1	50.7	33.6	36.5	1304
c	43.0	58.3	54.7	45.8	52.8	52.0	40.0	54.4	47.0	43.1	55.3	52.0	1624
b	3.5	12.8	13.1	6.2	4.8	6.3	19.6	24.6	21.3	5.8	10.5	11.1	291
a	0.0	0.0	0.0	0.0	0.3	0.0	2.7	3.6	2.6	0.3	0.5	0.4	15
	99.9	100.0	100.0	100.0	99.9	99.9	99.9	100.0	100.0	99.9	99.9	100.0	3234

TABLE 6.—COMPARISONS BETWEEN WEDGE FORMS AND RHOMBOHEDROIDS IN LITHOLOGY AND PER CENT ROUNDNESS, WITH INCREASING DISTANCE SOUTHWARD IN THE PLATEAU

Per cent calculated from data in Table 1. Localities grouped in zones as N (northern), M (middle), and S (southern)

	Limestone			Shale			Sandstone			Average		
	N	M	S	N	M	S	N	M	S	N	M	S
Wedge forms												
d	52.1	31.2	26.2	55.7	41.9	44.8	52.0	20.4	32.4	53.3	31.2	34.5
c	47.4	63.1	72.7	44.3	55.5	53.8	44.0	68.5	59.5	45.2	62.4	62.0
b	0.5	5.7	1.2	0.0	2.7	1.4	4.0	11.1	8.1	1.5	6.5	3.6
a	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0
	100.0	100.0	100.1	100.0	100.1	100.0	100.0	100.0	100.0	100.0	100.1	100.1
Rhombohedroids												
d	51.9	16.0	31.0	35.8	29.7	35.5	29.4	17.6	16.0	39.1	21.1	27.5
c	45.6	77.0	65.6	57.1	66.0	64.5	61.8	67.7	72.0	54.8	70.2	67.4
b	2.4	6.9	3.4	7.1	4.3	0.0	8.8	14.7	12.0	6.1	8.6	5.1
a	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0
	99.9	99.9	100.0	100.0	100.0	100.0	100.0	100.0	100.0	100.0	99.9	100.0

Rounding is believed to result chiefly from rotation during transport, and till-fabric interpretation indicates that wedge forms are less well adapted to sustained rotation than rhombohedroids (Holmes, 1941, p. 1339). With the long axis of a wedge form normal to the direction of forward movement, the intermediate axis or dimension is located near one end. Rotation would tend to advance this end faster than the pointed one and bring the long axis of the stone parallel with the direction of ice flow, in which position abrasion of its flat surfaces is more effective. By contrast, a rhombohedroid in rotation presents two similar ends to the impinging ice. Therefore it may roll forward evenly and is not likely to shift irregularly and rotate on either of its two lesser axes (although till-fabric studies show that some stones do this, especially if the three axes are of near-uniform lengths). Because a wedge form is much less stable in rotation, it is probably more likely to rotate erratically on a lesser axis, which would require a maximum vertical clearance between subjacent and superjacent stones also in transit. Crushing would therefore be more likely than with the more stable rhombohedroids. Confirmation of this inference is seen in the relatively greater decrease of *b* roundness wedge forms in passing from the middle zone to the southern.

Data discussed thus far are believed to show only that originally angular rhombohedroids and wedge forms undergo rounding by abrasion until crushing returns the fragments to the *d* state (and of course in smaller sizes). No trend toward shaping rhombohedroids into wedge forms is apparent.

Striation

Despite the relative rarity of well-striated stones in till, the expression "striated and faceted" became standard terminology long ago and has often been used unwisely.[1] In this study each stone was carefully examined for striae, with occasional use of a hand lens, but no record was made of extent or degree of striation. As Table 7 shows, about 28 per cent were striated; but most of these bore only slight abrasion marks. Striae are most easily seen on limestones because of their medium hardness, massive structure, and generally fine-grained texture, and because they take a high polish from transportation in a silty, clay-rich matrix. Most of the striated stones in the shale group are from the Genesee black shale, which is fine-textured and massive when freshly quarried. Sandstones are so hard that striae, if present, are generally too faint for clear recognition.

The fact that magnitude or depth of striae is roughly proportional to the size of the stone was observed long ago. Thus Miller (1884) wrote: "Many little boulders can be detected, measuring about the 1/100th part of an inch in length, bearing striae about the 1/2000th of an inch in diameter.[2]" These relationships show that stones move differentially in their transporting matrix, and that a graving point can impinge only as strongly as the stability of the stone in the inclosing ice will allow. Thus the larger stones receive the deeper striae, which probably accounts in part for the higher per cent striation reported for cobbles than for pebbles (Table 7).

On the more sharply angular stones the stronger striae tend to parallel one or another prominent edge, thereby indicating a controlled sliding of the stone and shearing movement of the inclosing ice. With slight rounding this tendency is much reduced. On well-rounded stones the individual striae traverse convex surfaces in apparently random orientation, although most of them can be traced through only a limited arc. These random striae are the least conspicuous of all and are believed to record relatively free movement in the transporting medium, which is presumed to be especially conducive to further rounding. Thus as long as the stone is in transit and escapes crushing, it becomes progressively less subject to "controlled" abrasion.

As implied in the preceding paragraph, most of the conspicuous striae on till stones occur on flat surfaces or in parallel sets tangential to the general convexity of the stone. This is evidence that they were made while the stone was not rotating. Either the stone was moved over the abrading materials which striated its under surface, or it was held in some medium while the abrading materials passed over its

[1] A true facet is produced by abrasion and is not merely a surface initially present, such as stratification or a fracture surface which has been striated later, although these initially flat surfaces, if abraded sufficiently, may not be readily distinguished from true facets.

[2] The term boulder was then unrestricted and designated merely a clastic fragment. This was the meaning when the term boulder clay came into use. Also the term boulder train was not intended originally to denote exclusively sizes exceeding 256 mm.

upper surface. Although both modes are probably effective under some circumstances, debate as to which is dominant began early and still continues. After careful observation, Miller (1884) favored the view that "the sole of the boulder, as Robert Chambers used to term their flatter side, seems generally, in fact, to have travelled uppermost." This implies abrasion by the faster moving superjacent ice, but the principle is equally applicable to abrasion while the stone is held stationary rather than being carried in a slower moving subjacent part of the glacier. The contrary opinion appears to rest on the assumption that stones at the base of the Pleistocene glaciers, even in the middle latitudes, were frozen with viselike solidity in the base of the glacier as it moved forward over an abrasive surface.

This latter concept is untenable. It implies an ice temperature far below the melting point, because ice near the melting point yields readily to pressure. The considerable amount of friction from abrasion and movement of ice already presumably at the pressure-melting point would produce a continuous supply of intergranular meltwater. Such ice would not hold a stone in constant position against a striating medium.

The only material believed to be of sufficient compactness to hold a stone in constant position while strong abrasion is in progress is the subglacially deposited till over which the ice continues to move its load of rock materials. This is essentially the mode of origin of striae on a boulder pavement. Such abrasion, if continued, would produce a true facet; but only by coincidence could the faceting develop a wedge-form stone.

TABLE 7.—STRIATED STONES CLASSIFIED AS TO SIZE AND LITHOLOGY, WITH INCREASING DISTANCE SOUTHWARD IN THE PLATEAU

Same data shown in both parts of table, arranged differently for easier comparisons. Percentage is based on the total number of stones in each of the several categories. Number of striated stones shown in parentheses. Locality groups or zones: N (northern), M (middle), S (southern)

	Limestone			Shale			Sandstone			Average
	N	M	S	N	M	S	N	M	S	N-M-S
Pebbles	42.4	42.7	57.1	10.1	13.0	11.5	0.7	1.3	..	26.9
Cobbles	56.0	63.0	54.5	16.7	31.4	43.8	..	10.0	..	43.5
Average	43.5	44.0	56.8	10.4	14.0	11.0	0.7	1.9	..	27.9

	N			M			S			Total average
	Limestone	Shale	Sandstone	Limestone	Shale	Sandstone	Limestone	Shale	Sandstone	
Pebbles	(336)	(14)	(1)	(234)	(79)	(2)	(115)	(32)	(0)	(813)
	42.4	10.1	0.7	42.7	13.0	1.3	56.5	11.5	..	26.9
Cobbles	(38)	(1)	(0)	(24)	(11)	(1)	(6)	(7)	(0)	(88)
	56.0	16.7	..	63.0	31.4	10.0	54.5	43.8	..	43.5
Total average	(374)	(15)	(1)	(258)	(90)	(3)	(121)	(39)	(0)	(901)
	43.5	10.4	0.7	44.0	14.0	1.9	56.8	11.0	..	27.9

Table 8 gives more detailed data on striation. Southward in the plateau, limestone rhombohedroids show appreciably higher per cent striation than almost any other limestone group. However, the striated rhombohedroids decrease in per cent southward while the others increase. The most notable increase is in the ovoid and discoid-tabular groups, so that in the southern zone the percentages are similar in nearly all the form groups. In the case of ovoids and discoids the increase may reflect the increasing coarseness of the till matrix, which produced recognizable striae, even though randomly oriented on these rounded forms, than would

have been the case had the associated matrix been more clayey.

The lower per cent of striation recorded for results in both crushing and abrasion of the stones transported. Much of the abrasion occurs on flat surfaces originally formed by stratifica-

TABLE 8.—STRIATED STONES CLASSIFIED AS TO LITHOLOGY, FORM, AND ROUNDNESS, WITH INCREASING DISTANCE SOUTHWARD IN THE PLATEAU

Number of striated stones shown in parentheses. Percentage is based on the total number of stones in each of the several categories. For example, the 87 striated stones constitute 80.5% of the entire number of limestone rhombohedroids in the northern zone. Locality groups or zones: N (northern), M (middle), S (southern)

	Limestone			Shale			Sandstone			Total average
	N	M	S	N	M	S	N	M	S	
Rhombohedroid	(87)	(63)	(18)	(5)	(24)	(8)	(0)	(1)	(0)	(206)
	80.5	48.0	62.0	11.9	14.8	10.2	..	2.9	..	27.7
Wedge-form	(190)	(106)	(42)	(1)	(43)	(10)	(0)	(0)	(0)	(393)
	45.2	40.3	50.0	2.0	16.4	6.9	..	..	..	28.6
Discoid-tabular	(29)	(49)	(29)	(8)	(14)	(12)	(1)	(0)	(0)	(142)
	35.4	49.5	66.0	32.0	11.2	22.6	4.0	..	..	28.1
Varihedroid	(59)	(16)	(13)	(1)	(6)	(2)	(0)	(1)	(0)	(98)
	46.1	36.4	46.4	7.7	8.7	3.8	..	8.3	..	26.3
Ovoid	(9)	(24)	(19)	(0)	(3)	(7)	(0)	(0)	(0)	(62)
	41.0	48.0	65.5	..	14.1	29.2	..	..	..	26.1
d	(163)	(42)	(33)	(2)	(35)	(6)	(0)	(1)	(0)	(282)
	35.5	24.7	13.2	2.9	13.9	4.1	..	3.4	..	21.6
c	(199)	(176)	(70)	(12)	(50)	(27)	(1)	(1)	(0)	(536)
	53.8	22.7	19.6	18.2	14.8	14.7	1.7	1.1	..	33.0
b	(12)	(40)	(18)	(1)	(5)	(6)	(0)	(1)	(0)	(83)
	40.0	27.2	23.7	11.1	16.2	26.1	..	2.4	..	28.5
a	(0)	(0)	(0)	(0)	(0)	(0)	(0)	(0)	(0)	(0)
	..	..	..	..	..	..	..	..	..	..

wedge forms than for rhombohedroids should be considered in the light of their total numbers, for wedge forms are nearly twice as abundant as rhombohedroids (Table 5). As shown in Table 8, the average per cent for each of the till-stone forms is remarkably uniform, ranging from 26.1 to 28.6 per cent.

CONCLUSIONS

This study confirms the interpretation of earlier investigators that glacial transport tion, jointing, or fracture, but a much greater amount is expressed in rounding of edges and corners. This rounding goes on until the stone is either crushed or deposited. On the crushed fragments, the rounding process begins anew. Faceting—the forming of an abraded flat surface where none existed before—is believed to be relatively rare. If crushing does not occur and if transportation continues, increasing roundness results finally in an ovoid form, or in a discoid form if the stone has tabular proportions. Neither field observation nor the

foregoing analysis indicates that long-continued glacial abrasion tends toward a wedge or flatiron form. The relatively prominent striae commonly observed on wedge-form stones

An incidental conclusion from this study is that intergranular meltwater arising from frictional heat of abrasion and glacier flow facilitates the downward migration of the fine

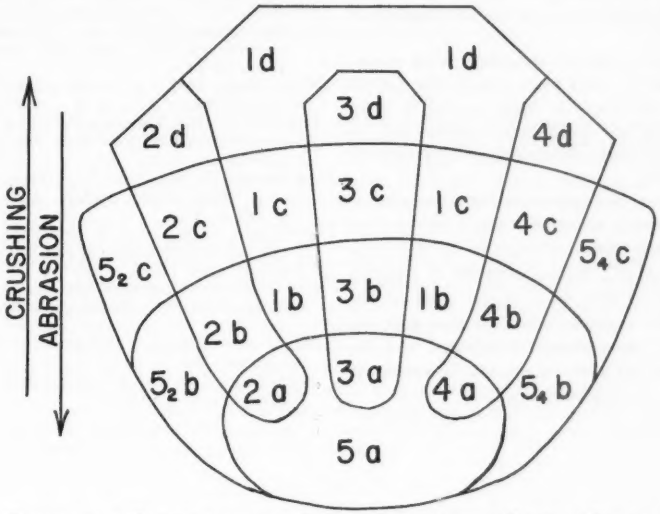

FIGURE 4.—DIAGRAMMATIC REPRESENTATION OF INFERRED RELATIONSHIPS BETWEEN SHAPES AND ROUNDNESS OF TILL STONES, AND PROCESSES AFFECTING THEM IN GLACIAL TRANSPORT

Roundness: *d* to *a*. Shapes: 1—Varihedroid (or heterohedroid) 2—Wedge-form (or cuneoid) 3—Tabular discoid 4—Rhombohedroid 5—Ovoid. Subscripts denote secondary resemblance.

seems rather to be a consequence of their original shape which is not conducive to rotation in transit. Because of this initial shape and its apparent greater susceptibility to sliding and therefore to striation, wedge forms may become rounded somewhat more slowly than others. All lithologic types show these progressive changes toward the ultimate ovoid form, but at rates inversely proportional to their resistance to abrasion and to crushing. In general, the larger stones are less susceptible to crushing than the smaller ones and therefore attain a higher average degree of roundness. Therefore von Engeln's secondary type, rounded but nonfaceted, is believed to be the primary type toward which glacial abrasion tends. Wedge forms in the central New York area are more abundant than stones of other shapes, and their striae are commonly more noticeable than those on the others. These factors, and the lack of careful distinction between true facets and mere striation of pre-existing surfaces, may account for the recognition they have received by earlier investigators.

products of abrasion and their deposition as the subglacial till matrix. This represents one phase of size sorting by a glacier, significantly different from that of flowing water in that the finer clastics are deposited more readily than the coarser.

Figure 4 shows qualitatively the essential relationships, inferred from this study, of form and roundness of till stones as affected by glacial transport. Most abrasion occurs as a rounding of edges and corners and tends ultimately to produce ovoid shapes. Faceting (as well as crushing) increases angularity but is omitted from the diagram because it is believed to be the exception rather than the rule.

Rounding of a varihedroid increases its symmetry, and beyond the *b* stage it is more properly referred to one of the other shapes. A stone of *d* roundness, which with a minimum of rounding would become an ovoid, is more properly classed in another group.

The term heterohedroid[3] may be substituted

[3] Suggested by Carl O. Dunbar.

for varihedroid, and cuneoid may replace wedge form if more uniformity of terms is desired. The designation tabular-discoid, rather than discoid-tabular, suggests more adequately the effects of rounding in this group. Stones referred to 3*a* and many in the 3*b* category are the discoids.

Subscripts to denote secondary form resemblance could be used more extensively than is indicated and may be found useful in more detailed analysis of till-stone shapes and response to the transportational and depositional environments.

In the New York plateau district, concentration of glacier movement along the valleys and the stony nature of much of the drift may have resulted in a ratio of abrasion to crushing different from that in a region such as the Midwest, for example, where forward flow was presumably more evenly distributed and the glacier load less stony. Comparative studies in other regions would be helpful.

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UPPER DEVONIAN CORRELATIONS IN WESTERN UTAH AND EASTERN NEVADA

By David L. Clark and Joseph H. Becker

Abstract

From the Devils Gate limestone, Pilot shale, and Pinyon Peak limestone of the eastern Great Basin, the writers have obtained 54 species of 28 conodont genera. Species of the genus *Palmatolepis* can be used in refined Devonian correlation and classification with the standard stages based on ammonoid genera in Europe. Three distinct assemblage zones, designated A, B, and C, are classified as Upper *Manticoceras*, uppermost *Manticoceras* or Lower *Cheiloceras* and Upper *Cheiloceras* or Lower *Platyclymenia*, respectively.

The upper part of the Devils Gate can be correlated with the lower and middle Pilot and the upper Pinyon Peak. Zone A, present in these formations, can be correlated with zone II of the Houy formation in Texas, the Independence shale and its equivalents in Iowa and Illinois, the Olentangy shale of Ohio, and the lower part of the Perrysburg formation and its equivalents in western New York and northern Pennsylvania.

Zones B and C, present in the middle-lower and middle part of the Pilot and upper part of the Devils Gate and Pinyon Peak, can be correlated with zone III of the Texas Houy formation, the Grassy Creek shale and its equivalents in Illinois and Iowa, and other sections in the mid-continent and eastern United States including the Huron member of the Ohio shale of Ohio and Kentucky and the middle and upper part of the Perrysburg formation in New York and Pennsylvania.

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Introduction

Certain Upper Devonian conodonts have long been noted as excellent index fossils, and recent studies in America (Youngquist, 1947), Europe (Bischoff and Ziegler, 1956), and Africa (Lys and Serre, 1957) have demonstrated the usefulness of certain species in correlation. Since the publication of conodont zonation of the German Devonian by Sannemann (1955a) and Müller (1956a) it appears possible to refine Upper Devonian correlation. Certain species of *Palmatolepis* have been demonstrated to be characteristic of each of the standard stages based on ammonoid genera (from older to younger, *Manticoceras*, *Cheiloceras*, *Platyclymenia*, *Oxyclymenia*, *Wocklumeria*), and therefore determination of rather precise correlations throughout the world appears possible on the basis of the more abundant microfossils.

One difficulty in establishing these refined

correlations in the American Upper Devonian is that no continuous section with representatives of each of the German stages is known, and though piecemeal sections have been composed they have obvious disadvantages (Müller and Müller, 1957, p. 1069–1070):

"Unfortunately, the localities for previous studies in North America expose only one or, at most, a few conodont-bearing horizons. Also, these localities are too far apart to be correlated exactly by physical means. Therefore, few authors have been in a position to realize fully the possibilities for a subdivision of the Upper Devonian on the basis of conodonts . . . . Such a section may be present somewhere in the Rocky Mountains, where little conodont work has yet been done."

During the summer of 1950, Walter Youngquist (1958, personal communication) obtained a conodont fauna in the Confusion Range of western Utah from rocks associated with *Manticoceras*, the form genus of the lowermost of the five Upper Devonian stages. Chapman (1958, M.S. thesis, Univ. Kansas) recently described this fauna. In 1957, James E. Brooks sampled the Upper Devonian Pilot shale at the same locality. Certain of Brooks' samples contained conodonts indicative of the second or *Cheiloceras* stage. On the basis of the occurrence of conodonts representing the lower two of the five Upper Devonian stages, the writers were impressed with the possibility of refining Upper Devonian correlations in the Cordilleran and possibly finding a more nearly continuous section representing standard stages than had been described in North America. Complete sections of the Pilot shale and its equivalents from localities in central Nevada to central Utah were measured and sampled with this in mind (Fig. 1).

Acknowledgments

This study was aided by a grant from the Penrose Bequest of The Geological Society of America. Dr. J. E. Brooks, Southern Methodist University, provided samples from one section and contributed to stratigraphic determinations. Mr. Lawrence Lenzen, Jr., aided as field assistant, and Dr. K. J. Müller of the Technische Universität Berlin verified identifications. Figured specimens are in the Paleontology Repository, Brigham Young University (B.Y.U.).

Procedure

Before the collection of field samples for this study, a detailed investigation was made in the laboratory to discover the type of sediment which yielded the best preserved and most abundant conodonts. Twenty-five samples from many parts of North America and from Cambrian to Jurassic in age were studied and the following recorded: color, grain size, amount of effervescence in 15 per cent HCl, associated fossils, bedding characteristics, and primary and secondary structures. Following this descriptive phase all samples were crushed to pieces about half an inch in diameter, and 250 grams were used for preliminary examination. Limestones and other calcareous types were dissolved in 15 per cent glacial acetic acid, and the more siliceous fragments were boiled in a 20 per cent solution of NaOH (Bolli, 1950).

The residues were washed and sieved, and material retained on the 60 mesh (0.250 mm) and 120 mesh (0.125 mm) was examined. Conodonts were found only in the calcareous sediment which contained other fossils. About 80 per cent of the conodont-bearing rocks was fine- to medium-grained bioclastic limestones; cephalopods were the associated fossil in 60 per cent of the rocks and gastropods in 40 per cent.

These observations were made on a small selective sample and do not represent the only environmental types favorable for conodonts; however, conclusions can be drawn as to which rocks are more favorable.

Stratigraphy

Introduction

The Devils Gate limestone, Pilot shale, and Pinyon Peak limestone were measured and sampled at various localities in central Nevada to central Utah. Certain samples obtained were treated in the field in the same manner as those samples checked before field work in the laboratory. Those localities from which conodont-bearing samples were obtained were revisited and critical beds traced and sampled extensively.

Devils Gate Limestone

Merriam (1940, p. 58) defined the Devils Gate limestone as the upper portion of the Nevada limestone of Hague (1892) and chose the north side of Devils Gate Pass on the Lincoln Highway about 8 miles northwest of Eureka, Nevada, as the type section.

The formation consists of medium-gray to black, thick-bedded limestone and thin inter-

bedded shaly limestone. The thickness ranges from 675 feet in the Diamond Range (Nolan *et al.*, 1956, p. 49) to 2065 feet where partial sections at Devils Gate Pass and Modoc Peak interbedded with siltstone and shale at the base, which some students have regarded as uppermost Guilmette (Chapman, 1958, M.S. thesis, Univ. Kansas). The contact with the

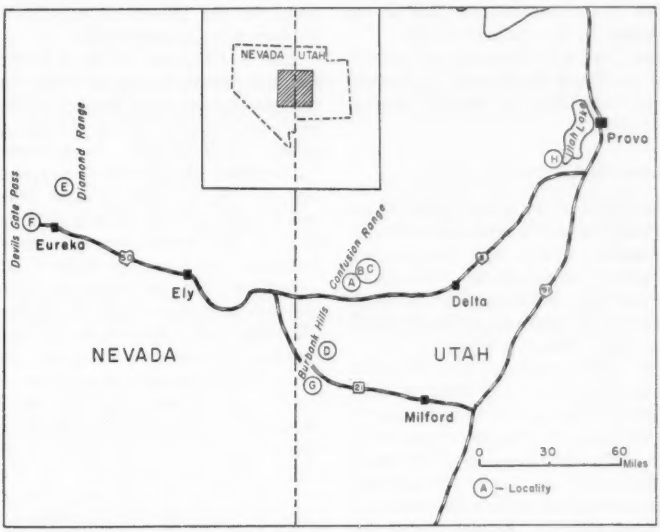

FIGURE 1.—INDEX AND LOCALITY MAP OF UPPER DEVONIAN SECTIONS IN WESTERN UTAH AND EASTERN NEVADA

in central Nevada were combined (Merriam, 1940, p. 16). Merriam (p. 58) considered it of Middle and Late Devonian age.

Samples of the Devils Gate were obtained from the type section on U. S. Highway 50 and from a section on the west side of the Diamond Range, north of Eureka, Nevada.

Pilot Shale

The Pilot shale was named by Spencer (1917, p. 24) for a section exposed on Pilot Knob near Ely, Nevada. It is equivalent to the lowest beds of the White Pine shale of Hague (1892, p. 65). Nolan *et al.* (1956, p. 52) report that the thickness of the Pilot ranges from 315 to 1000 feet at localities in eastern Nevada. The formation consists of platy calcareous shale with interbedded limestone and siltstone. The limestones are thicker and arenaceous in the lower part, and the contact with the underlying Guilmette is gradational. The writers have followed Brooks (1954, Ph.D. thesis, Univ. Washington) in considering the contact to be at the top of the last massive limestone of the Guilmette. Thus defined, the Pilot includes some thick-bedded limestones overlying massive limestones of the Joanna is well defined at most localities. Nolan *et al.* (1956, p. 53) suggested that the upper Pilot is of Early Mississippian age, but evidence is lacking.

Sections of the Pilot were measured and sampled in the Diamond Range, Nevada, and in the Confusion Mountains and Burbank Mountains of western Utah (Fig. 2). In one area of good outcrops in the Confusion Range a single arenaceous limestone bed was sampled at closely spaced intervals over a distance of 8 miles. Two sections were measured in different parts of the Burbank Mountains, north and south of Utah State Highway 21, about 17 miles east of Garrison, Utah.

Pinyon Peak Limestone

Lindgren and Loughlin (1919, p. 36) named the Pinyon Peak limestone for exposures in the Tintic district of west-central Utah. This unit is overlain conformably by the Mississippian Gardner dolomite and underlain with apparent conformity by Devonian Victoria quartzite. In central Utah the thickness ranges from 73 to 300 feet (Peterson, 1956). The

Pinyon Peak consists of gray, thin- to medium-bedded limestones and dolomites and has been called Late Devonian on the basis of meager fossil evidence and its position below Early Mississippian rocks (Clark, 1954, p. 6). Rigby (1959) described tectonic relationships of the Upper Devonian in this part of Utah.

The Pinyon Peak was measured and sampled northeast of the Tintic district; on Rattlesnake Spur west of Utah Lake, a compete section was obtained (Fig. 2).

Conodont Assemblage Zones

Conspecific forms of the genus *Palmatolepis* have been found in widely separated parts of the world (Müller, 1956b), and the presence of certain species in association with stratigraphically restricted ammonoids in Germany has been reported (Sannemann, 1955a; Bischoff and Ziegler, 1956; Müller, 1956a).

More than 5000 conodonts were obtained from samples taken from 113 different beds at 31 localities in Utah and Nevada. About 60 per cent of the fauna is made up of species of *Palmatolepis*. Although 54 species of 28 genera were identified, only species of *Palmatolepis* were used in correlations as they are evidently the most "short-ranged." Rhodes (1952) described natural conodont assemblages which include a number of bars and blades but only a pair of platform types such as *Palmatolepis*. Many of the bar and blade types range from Ordovician to Triassic, and speciation has been so extensive that species and even generic affinities are extremely difficult to determine. Until monographic treatment of many of these genera is made, identifications are so subjective that these conodont types are of little stratigraphic value.

Platform types are characterized by short stratigraphic ranges, and morphological "lineages" can be constructed (Müller, 1956a, p. 33–36). In precise zonation, the writers agree with Scott (1958, Ph.D. thesis, Univ. Illinois) who dealt almost exclusively with the platform types in zonation of the Upper Devonian and type Mississippian. The 44 species associated with *Palmatolepis* in this study will be described in a separate report.

Three distinct assemblage zones based on species of *Palmatolepis* can be recognized in the Cordilleran Upper Devonian: zone A, representing Upper *Manticoceras* strata, zone B, characterized by species common to uppermost *Manticoceras* and Lower *Cheiloceras* strata, and zone C, representing Upper *Cheiloceras* or Lower *Platyclymenia* strata. Zones A and B have one species of *Palmatolepis* in common as do zones B and C, but no one species occurs in all zones. Although probably more overlapping of species occurred than was found in the present study, the distinction between zones is remarkable.

ZONE A: Chapman (1958, M.S. thesis, Univ. Kansas) described a small fauna representing this assemblage zone. Species of *Palmatolepis* which are characteristic of this zone are *Palmatolepis* (*Palmatolepis*) *linguiformis* Müller, *Palmatolepis* (*Manticolepis*) *marginata* (Stauffer) var. A, *P.* (*M.*) *foliacea* (Youngquist), and *P.* (*M.*) *subrecta* (Miller and Youngquist). Only *P.* (*P.*) *linguiformis* is common to both European and American rocks of Late *Manticoceras* age, but the other species have been found in association with *Manticoceras* in the Olentangy and Amana shales in the mid-continent. Chapman's collection (1958, M.S. thesis, Univ. Kansas) came from rocks associated with *Manticoceras* in the Confusion Mountains. The writers obtained their collection of specimens from rocks about 50 feet above the last occurrence of *Manticoceras* and from 50 to 150 feet above the base of the Pilot in the Confusion Mountains and from about 700 feet above the base of the Devils Gate section at Devils Gate Pass. At the latter locality representatives of zone A were found through 200 feet of section.

The fauna reported by Hass (*in* Nolan *et al.*, 1956, p. 53) from the lower beds of the Pilot on the east side of Newark Mountain, Nevada, belongs to this zone. The occurrence of conodonts in association with *Cyrtospirifer* in the upper beds of the Devils Gate at Devils Gate Pass gives evidence of the correlation between these beds and the basal beds of the Pilot shale as suggested by Nolan *et al.* (1956, p. 51):

> "A conodont fauna such as occurs in the lower part of the Pilot Shale has not yet been found in association with *Cyrtospirifer* at Devils Gate; if one should be found, it would provide needed proof of this suggested correlation."

This correlation also substantiates the idea that the decreased thickness of the Devils Gate to the east of the type section is partially "due to replacement of the higher beds (of the Devils Gate) . . . by the Pilot Shale eastward" (Nolan *et al.*, 1956, p. 49).

This zone, classified as Upper *Manticoceras*, is widespread, and its representatives are the most numerous of species found. The only

FIGURE 2.—STRATIGRAPHIC SECTIONS OF THE PILOT SHALE, DEVILS GATE LIMESTONE, AND PINYON PEAK LIMESTONE IN WESTERN UTAH AND EASTERN NEVADA

Assemblage zones and classifications are indicated. Numbers indicate sample numbers

locality where diagnostic elements of this zone were not found was in the Burbank Mountains, and this was probably due to insufficient sampling.

ZONE B: This zone is characterized by *Palmatolepis* (*Manticolepis*) *marginata* (Stauffer) var. B, *P.* (*M.*?) *quadrantinodosalobata* (Sannemann), *P.* (*Palmatolepis*) *subperlobata* Branson and Mehl, *P.* (*Deflectolepis*) *coronata* Müller, and *P.* (*D.*) *minuta* (Branson and Mehl). The subgeneric names including the status of *Deflectolepis* in light of the recent criticism by Hass (1959) are discussed below under Systematic Paleontology.

This zone occurs 100–200 feet above zone A. The occurrence of *P.* (*M.*) *marginata* in both zones indicates the close relationship of the two faunas. However, specimens of *P.* (*M.*) *marginata* occurring in zone B can be readily distinguished from the zone A specimens. The reduction in size of plate and other features which can be seen in evolutionary sequences of *Palmatolepis* (Müller, 1956a, p. 33–36) have led the writers to suggest varietal difference for the specimens in the two zones: var. A in zone A, and var. B in zone B.

Of the *Palmatolepis* species found in this zone all except the uniquely American *P.* (*M.*) *marginata* have been reported in Europe. *P.* (*M.*?) *quadrantinodosalobata* is seemingly restricted to uppermost *Manticoceras* stufe and Lower *Cheiloceras*, *P.* (*P.*) *subperlobata*, throughout the *Manticoceras* into the Lower *Cheiloceras* beds, *P.* (*D.*) *coronata*, to uppermost *Manticoceras* and lowermost *Cheiloceras*, *P.* (*D.*) *minuta*, to Upper *Manticoceras* through *Platyclymenia* (III and IV).

It is not possible to determine whether this zone is Upper *Manticoceras* or Lower *Cheiloceras* on species of *Palmatolepis*. Müller (1956a, p. 35) records *P.* (*D.*) *coronata* only in uppermost *Manticoceras* stufe and *P.* (*D.*) *minuta* from Lower *Cheiloceras* through *Platyclymenia* stufe. The two species occur together in zone B, which may indicate that Sannemann (1955a, p. 326) is correct in assuming that *P.* (*D.*) *minuta* occurs first in Upper *Manticoceras* stufe and may further indicate that zone B is of very Late *Manticoceras* age. However, in the Great Basin fauna there is an almost complete morphological range from *minuta* to *coronata*, and all other zone B species of *Palmatolepis* are known from *Cheiloceras* age rock. The fact that zone B species occur several hundred feet above definite Upper *Manticoceras* species of zone A supports the idea that zone B represents uppermost *Manticoceras* or Lower *Cheiloceras* strata.

ZONE C: The youngest zone determined in the eastern Great Basin Devonian is characterized by three species of *Palmatolepis*: *P.* (*P.*) *distorta* Branson and Mehl, *P.* (*P.*) sp. aff. *crepida* Sannemann, and *P.* (*D.*) *minuta* (Branson and Mehl). Only *minuta* is present in both zones B and C. It ranges from Upper *Manticoceras* through *Platyclymenia* strata. *P.* (*P.*) *distorta* is restricted to Upper *Cheiloceras* through *Oxyclymenia* strata according to Müller (1956a, p. 35) although Sannemann (1955a, p. 326) records its highest occurrence as lowermost *Platyclymenia* (IIIa). *P.* (*P.*) *crepida* has only been reported from Middle *Cheiloceras* strata in Germany. The fact that it occurs with species never found lower than Upper *Cheiloceras* in zone C suggests slightly different ranges in this country. However, the writers' specimens show some difference from *crepida* s. s., as described below, and may represent a new species. Therefore, zone C represents at least Upper *Cheiloceras* age strata and probably Lower *Platyclymenia*.

This zone was recognized from the Diamond Mountains in Nevada to the Burbank and Confusion mountains in Utah. In the Diamond Mountains near the Phillipsburg mine the highest bed of the Devils Gate limestone, which is overlain by Pilot shale, contains elements of this zone. To the east in western Utah, this zone occurs about 350 feet above the base of the Pilot shale. This substantiates the correlation of uppermost Devils Gate with middle Pilot from central Nevada to western Utah.

At one locality in the Burbank Mountains the Pilot is 500 feet thick, and zone C occurs 283 feet below the overlying Lower Mississippian Joanna limestone. The Pilot at this locality has been considered Mississippian by Rush (1951, p. 14–18). In the Confusion Range where the Pilot is 1100 feet thick only fragments of specimens from this zone were encountered 120 feet above zone B or about 750 feet below the Joanna limestone.

This zone contains fossils representing the youngest Devonian that is presently recognized in the eastern Great Basin. The writers assume that the continuous nature of the Pilot above zone C indicates that representatives of the younger Upper Devonian stages may be present, and with their identification the Devonian-Mississippian boundary problem may be clarified. Lower Mississippian conodont species have not been described from the

Great Basin; in the course of this investigation they were noted at several localities, but their relationship to zone C is uncertain. Although the boundary may be in the upper Pilot shale as is often suggested (Brooks, 1954, Ph.D. thesis Univ. Washington; Nolan *et al.*, 1956) the continuous nature of the Pilot above zone C and the stratigraphic distribution of the zones below suggests that the Devonian Mississippian boundary is closer to the physical Pilot-Joanna boundary than heretofore suggested. Work in progress may substantiate this.

Correlations

Few studies of Upper Devonian conodonts in this country have been made on sections where multiple conodont zones were encountered. This is due to the rather noncontinuous nature of most Devonian outcrops as well as the unrefined work. The recent papers by Hass (1947; 1958) and Müller and Müller (1957) concern several lower Upper Devonian zones, but before exact correlation can be made with most sections studied, more refined work is needed.

On the basis of the three zones defined, correlations can be effected within the eastern Great Basin as well as with other sections in North America, and these zones are classified with the European standard. The occurrence of species indicative of zone A in the *Cyrtospirifer* or uppermost zone of the Devils Gate limestone (Merriam, 1940, p. 61) and in the lower 200 feet of the Pilot shale in western Utah substantiates this correlation (Nolan *et al.*, 1956, p. 51). Zone A of these formations can also be correlated with the fauna of zone II of the Houy formation in central Texas (Cloud *et al.*, 1957, p. 812), the Dowelltown member of the Chattanooga shale of central Tennessee (Hass, 1956), the Independence shale, Lime Creek and part of the Grassy Creek of Iowa (Müller and Müller, 1957), the Sweetland Creek shale of Iowa (Miller and Youngquist, 1947), the Sylamore sandstone and equivalents of western Illinois (Scott, 1958, Ph.D. thesis, Univ. Illinois, *Palmatolepis* (*M.*)-*Ancyroides* zone), the lower beds of the Blackiston and Blocher formations of Indiana (Campbell, 1946), the Olentangy shale of Ohio (Stauffer, 1938), middle part of Arkansas novaculite (Hass, 1951), and the lowermost part of the Perrysburg formation of western New York and northern Pennsylvania (Hass, 1958). This zone can be classified with the Upper *Manticoceras* stufe of the European standard.

Zone B cannot be precisely correlated with any conodont assemblage previously described in North America because previous descriptions lack the refinement that this zone represents. It is correlated with the Sheffield shale and Gassaway member of the Chattanooga shale (Müller and Müller, 1957, p. 1077) and is equivalent in part to the Cassadaga stage of Cooper (Hass, 1958, p. 767). This zone can be classified as uppermost *Manticoceras* or Lower *Cheiloceras*.

The writers did not find zone B in the Devils Gate limestone but traced it extensively in the Pilot shale of western Utah (Fig. 3).

Zone C was recognized in the uppermost bed of the Devils Gate limestone in the Diamond Mountains and in the middle and upper-middle part of the Pilot shale in the Burbank and Confusion mountains. In addition to the *Palmatolepis* fauna, zone C in the Diamond Mountains is characterized by species of the conodont genus *Polygnathus* which are abundant in the upper part of the Pinyon Peak limestone on Rattlesnake Spur in central Utah. The Pinyon Peak contains a few species similar to those found in zone A, but the species of *Polygnathus* are those which are common in zone C age rock in other parts of the country; although the evidence is inconclusive, the writers tentatively correlate the upper Pinyon Peak with zone C.[1]

Species of *Palmatolepis* in zone C can be correlated with species in zone III of the Houy formation in central Texas (Cloud *et al.*, 1957), part of the Gassaway member of the Chattanooga shale in central Tennessee (Hass, 1956), the middle and upper Grassy Creek shale of Illinois (Scott, 1958, Ph.D. thesis, Univ. Illinois, *Palmatolepis* (*P.*)-*Polylophodonta* zone), the upper part of the lower Blackiston formation of southern Indiana (Campbell, 1946), the Huron member of the Ohio shale of Ohio and Kentucky (Hass, 1947; 1958), and all except the lowermost portion of the Perrysburg formation of western New York and northeastern Pennsylvania (Hass, 1958). This zone is classified as Upper *Cheiloceras* or Lower *Platyclymenia* of the European standard.

Upper Devonian stratigraphically higher than that represented by zone C is not well

[1] While this paper was in press, the senior writer found species of *Palmatolepis*—e.g., *P.* (*Palmatolepis*) *rugosa*—and other genera in the Upper Pinyon Peak which confirm this correlation.

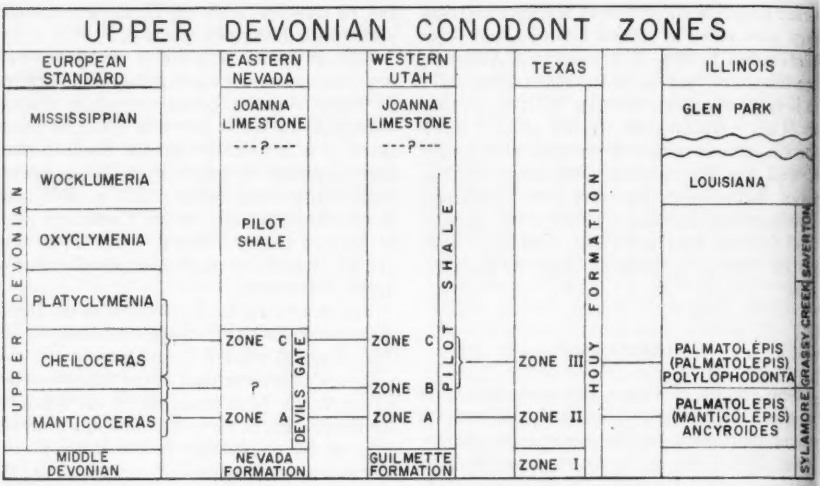

FIGURE 3.—UPPER DEVONIAN CONODONT ZONES AND CORRELATIONS

known in this country. Scott's determination of *Platyclymenia* and younger age for his *Palmatolepis* (*P.*)-*Polylophodonta* zone in Illinois and Iowa is not well substantiated (1958, Ph.D. thesis, Univ. Illinois). Klapper (1958) has described a small fauna from the Darby formation of the Wind River Mountains, Wyoming, which he considers as *Oxyclymenia* (to V) of the European standard based on the occurrence of *P.* (*P.*) *gonioclymenia*. Nothing this young has been found in the Great Basin Devonian. *Wocklumeria* age strata (the youngest Upper Devonian) may not be distinguishable in North America on the basis of conodonts alone, as only two species of *Palmatolepis* are known from this age rock, and they are present in older strata as well.

SYSTEMATIC PALEONTOLOGY

Genus *Palmatolepis* Ulrich and Bassler, 1926

Type species: *P. perlobata* Ulrich and Bassler

On the basis of evolutionary series of species of *Palmatolepis*, Müller (1956a) proposed the recognition of three subgenera, *P.* (*Palmatolepis*), *P.* (*Manticolepis*), and *P.* (*Deflectolepis*). Because of the clearly defined morphological and stratigraphic significance of these subgenera students who have never worked with conodonts can easily distinguish the subgenera and make valid stratigraphic determinations based on such identifications.

Hass (1959) commented on the apparent confusion between *Palmatolepis* and *Panderodella*. The genus *Panderodella* was proposed by Bassler (1925) with type species *Panderodella truncata* mentioned by Bassler (1925) and described by

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PLATE 1.—SPECIES OF THE GENUS *PALMATOLEPIS*

Figures 1–14 ×35; Figure 15 ×25

FIGURE 1–5—*Palmatolepis* (*Deflectolepis*) *coronata* Müller, all from section C and from different samples laterally along bed 19, B. Y. U. 525, 526, 627, 528 (Figures 4 and 5 upper and lateral views)

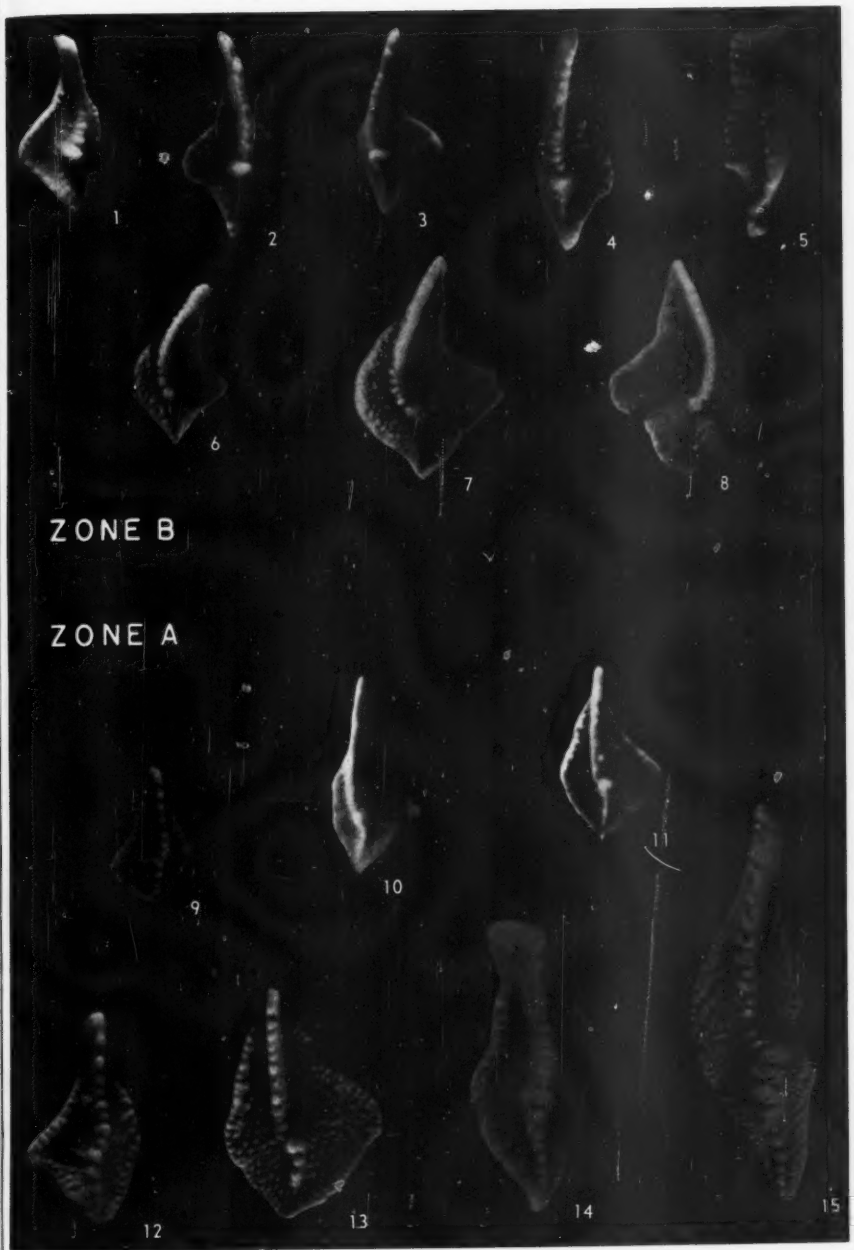
FIGURE 6–8.—*Palmatolepis* (*Manticolepis?*) *quadrantinodosalobata* (Sannemann), all from section C and from different samples laterally along bed 19, B. Y. U. 529, 530, 531

FIGURE 9–11.—*Palmatolepis* (*Manticolepis*) *marginata* (Stauffer) var. A, all from section B, bed 34, B. Y. U. 532, 533, 534

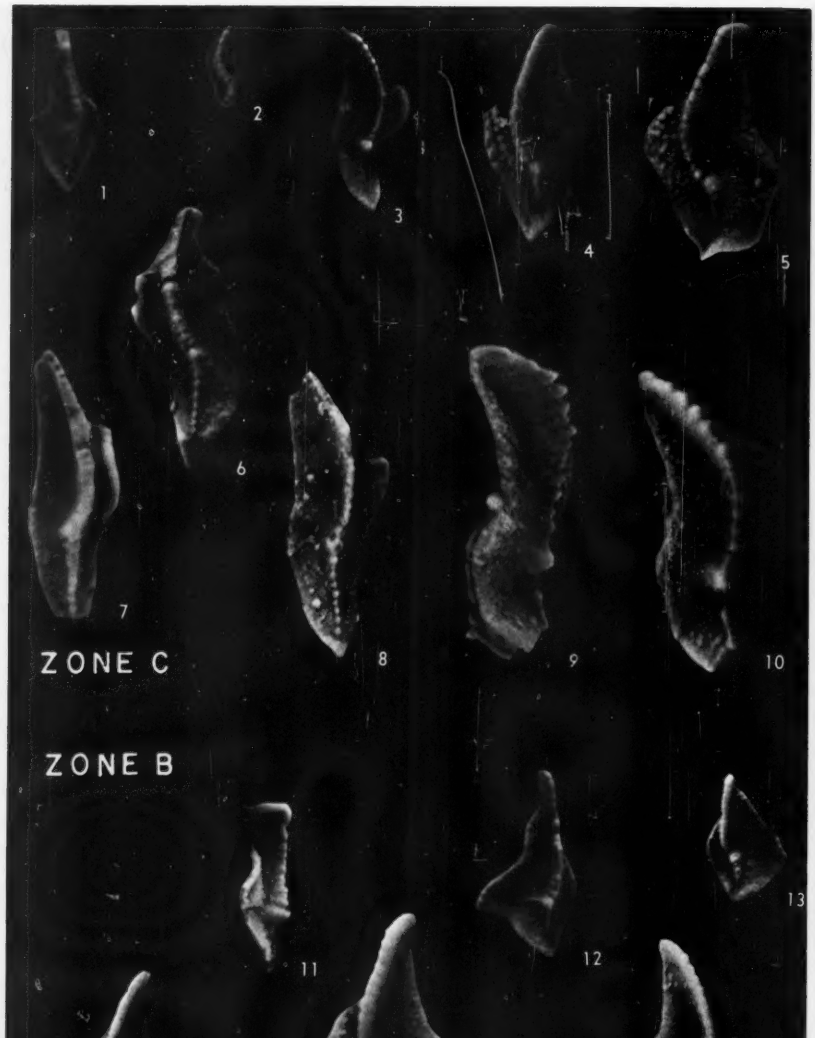
FIGURE 12.—*Palmatolepis* (*Manticolepis*) *subrecta* (Miller and Youngquist), section B, bed 34, B. Y. U. 535

FIGURE 13.—*Palmatolepis* (*Manticolepis*) *foliacea* (Youngquist), section B, bed 34, B. Y. U. 536

FIGURE 14–15.—*Palmatolepis* (*Palmatolepis*) *linguiformis* Müller, 14, from the Lower Blackiston formation, southeast of Sellersburg, Indiana, B. Y. U. 537, 15, from section B, bed 34, B. Y. U. 538



SPECIES OF THE GENUS *PALMATOLEPIS*



SPECIES OF THE GENUS *PALMATOLEPIS*

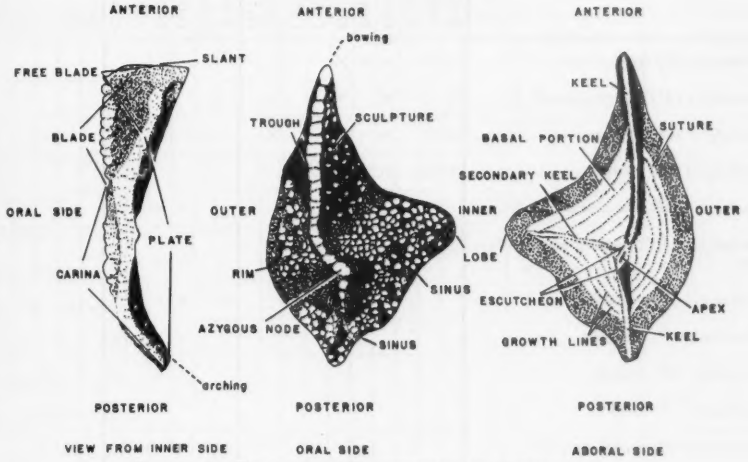


FIGURE 4.—DESCRIPTIVE TERMINOLOGY OF *PALMATOLEPIS*
(After Müller and Müller, 1957)

Ulrich and Bassler who illustrated three cotypes (1926, Pl. 9, figs. 15–17). Müller (1956a, p. 16) designated the specimen illustrated in Ulrich and Bassler's Figure 15 as lectotype and considered it a junior synonym of *Polygnathus* Hinde 1879. Hass (1959, p. 369) claims, however, that the specimen of Figure 15 is not *Polygnathus* but is conspecific with the specimen described as *Palmatolepis glabra* Ulrich and Bassler (1926); he lists *P. glabra* as a junior synonym of *Panderodella*. If Hass' determination is valid the specimen described as *Palmatolepis glabra* = *Panderodella truncata*, not *Panderodella glabra* as he states (1959, p. 369); *Palmatolepis* would become a junior synonym of *Panderodella*. Hass recognized that the well-known name *Palmatolepis* should be preserved. This could probably best be effected by regarding the fragments that *Panderodella* were based on as *nomen dubium*. Perhaps other specialists should also study the type material about which Müller (1956a) and Hass (1959) cannot agree.

Subgenus *Palmatolepis* (*Palmatolepis*)
Ulrich and Bassler, 1926

Type species: *P. perlobata* Ulrich and Bassler

Most species of this subgenus are distinguished by the fact that in all ontogenetic stages the posterior tip of the specimen is pointed upward. Müller (1956a, p. 24) has described exceptions. In some cases the sculpture is weaker than in other subgenera, and secondary carina and keel may be present (Fig. 4). In Europe most of these species occur in *Cheiloceras* and younger beds.

Palmatolepis (*Palmatolepis*) *distorta*
Branson and Mehl

(Pl. 2, figs. 2, 6–8)

Palmatolepis distorta BRANSON AND MEHL, 1934, Univ. Missouri Studies, v. 8, p. 237, Pl. 18, figs.

PLATE 2.—SPECIES OF THE GENUS *PALMATOLEPIS*

×35

FIGURE 1.—*Palmatolepis* (*Deflectolepis*) *minuta* (Branson and Mehl), section E, bed 1, B. Y. U. 518

FIGURE 2, 6–8.—*Palmatolepis* (*Palmatolepis*) *distorta* Branson and Mehl, all from section E, bed 1, B. Y. U. 514, 521, 517, 523

FIGURE 3–5, 9, 10.—*Palmatolepis* (*Palmatolepis*) sp. aff. *crepida* Sannemann, 3, from section C, bed 12, others from section E, bed 1, B. Y. U. 522, 513, 515, 516 (Figures 9 and 10, lateral and upper views)

FIGURE 11.—*Palmatolepis* (*Deflectolepis*) *coronata* Müller, section C, bed 19, B. Y. U. 525, lateral view

FIGURE 12, 13.—*Palmatolepis* (*Manticolepis*) *marginata* (Stauffer) var. B, section C, bed 19, B. Y. U. 519, 524

FIGURE 14–16.—*Palmatolepis* (*Palmatolepis*) *subperlobata* Branson and Mehl, section C, bed 19, B. Y. U. 511, 512, 520

TABLE 1.—DISTRIBUTION OF UPPER DEVONIAN CONODONT GENERA*

	ZONE A										ZONE B						ZONE C						
	B-34	B-34-2	B-34-3	B-34-12	B-34-15	F-5	F-6	F-9	F-15	F-17	B-26-A	B-26-B	C-19	C-19-11	C-19-18	C-19-21	G-6	G-9	E-1	C-12	A-15	H-1	H-3
Palmatolepis (P) linguiformis	X	X		X	X	X																	
Palmatolepis (M) marginata var. A	X	X	X	X		X			X														
Palmatolepis (M) foliacea	X	X		X		X			X														
Palmatolepis (M) subrecta	X	X		X		X																	
Palmatolepis (M) marginata var. B													X	X		X							
Palmatolepis (M?) quadrantinodosalobata													X	X	X	X							
Palmatolepis (P) subperlobata											X		X	X	X	X							
Palmatolepis (D) coronata													X	X		X							
Palmatolepis (D) minuta													X	X		X			X				
Palmatolepis (P) distorta																	X	X					
Palmatolepis (P) sp. aff. *crepida*																	X	X					
Ancyrodella n. sp.		X		X																			
Ancyropenta n. sp.				X			X																
Bryantodus sp.		X		X							X			X		X							
Centognathodus sp. A													X										
Centognathodus sp. B														X									
Ctenopolygnathus brevilamina	X																						
Ctenopolygnathus iowaensis						X																	
Ctenopolygnathus sp.														X							X		
Ellisonia plana																X							
Ellisonia sp.														X	X								
Euprioniodina gracilis													X										
Hibbardella sp. A				X									X	X									
Hibbardella sp. B				X		X		X		X			X	X		X							
Icriodus alternatus	X						X									X							
Icriodus curvatus																			X				
Icriodus rectus													X	X									
Icriodus sp.				X		X										X							
Ligonodina sp.				X		X	X					X	X				X			X			
Lonchodina sp.				X	X	X				X			X	X		X							
Lonchodus sp.					X									X			X			X			
Nothognathella abnormis		X			X																		
Nothognathella delawarensis											X		X	X									
Nothognathella n.sp.				X									X	X		X							

* Sample numbers refer to Figure 2.

TABLE 1.—*Continued*

	ZONE A										ZONE B						ZONE C						
	B-34	B-34-2	B-34-3	B-34-12	B-34-15	F-5	F-6	F-9	F-15	F-17	B-26-A	B-26-B	C-19	C-19-11	C-19-18	C-19-21	G-6	G-9	E-1	C-12	A-15	H-1	H-3
Nothognathella sp.		X		X															X				
Ozarkodina sp.						X	X			X			X	X									
Pdekysgnathus planus													X										
Pdekysgnathus n.sp.														X		X							
Polygnathus foliata				X									X	X		X							
Polygnathus linguiformis				X															X			X	
Polygnathus longipostica				X						X													
Polygnathus nodocostata																			X	X			X
Polygnathus normalis				X																			X
Polygnathus webbi				X				X					X	X	X	X						X	
Polygnathus sp.			X	X			X							X		X						X	
Prioniodella sp.													X										
Prioniodina sp.													X							X			
Prinoiodus sp.				X																			
Roundya sp.				X																			
Scutula sp. aff. *bipennata*													X										
Synprioniodina sp.												X		X									
Subbryantodus sp.													X										
Tripodellus n. sp. A													X			X							

13, 14; MÜLLER, 1956a, Senck. Naturf. Gesell. Abh. 494, p. 26, Pl. 7, fig. 20, 21

Plate thick, long, narrow, and sigmoidal. Inner side of plate extends full length of blade, outer part of plate meets blade at about mid-length between anterior tip and azygous node. Carina extends to posterior tip, azygous node same size as carina nodes or larger. Oral (upper) surface smooth. Narrow side of plate folded into ridge which is long and higher than carina and separated from it by deep narrow groove.

Remarks.—The sigmoidal shape of this species which aids in distinguishing it from *P.* (*P.*) *glabra* is variable. The writers have a complete series of strongly sigmoidal to only slightly curved forms, all of which have the high, long outer ridge and deep groove characteristic of *P.* (*P.*) *distorta*.

Occurrence.—Zone C (Table 1)

Palmatolepis (*Palmatolepis*) sp. aff. *crepida* Sannemann

(Pl. 2, figs. 3–5, 9, 10)

Palmatolepis crepida SANNEMANN, 1955b, Senck. Lethaea, v. 36, p. 134, Pl. 6, fig. 21

Plate thick, long, subovate; posterior of azygous node strongly flexed upward. Heavy nodose sculpture concentrated on anterior and middle part of plate. Carina and blade strongly sigmoidal, outer part of plate arched and rounded. A small inner lobe occurs on some specimens.

Remarks.—This differs from *crepida* s. s. in the very strongly upward flexed posterior end and in the tendency for an inner lateral lobe to develop. The stratigraphic significance of this has been discussed under zone C.

Occurrence.—Zone C (Table 1)

Palmatolepis (*Palmatolepis*) *linguiformis* Müller

(Pl. 1, figs. 14, 15)

Palmatolepis (*Palmatolepis*) *linguiformis* MÜLLER, 1956a, Senck. Naturf. Gesell. Abh. 494, p. 24, Pl. 7, figs. 1–7

Large heavy plate with very high stout blade, slender and sculptured with low ridges and nodes, slight tendency for development of inner lateral lobe. Posterior end flexed upward, tip may curve slightly downward. Blade slightly curved and extremely high at anterior end at which point it is free of narrow plate. Azygous node may not be distinct from carina which extends to posterior end.

Occurrence.—Zone A (Table 1)

Palmatolepis (*Palmatolepis*) *subperlobata* Branson and Mehl

(Pl. 2, figs. 14–16)

Palmatolepis subperlobata BRANSON AND MEHL, 1934, Univ. Missouri Studies, v. 8, p. 235, Pl. 18, figs. 11, 21

Large thin plate with well-developed inner lateral lobe opposite or slightly anterior of azygous node. Faint secondary carina of one or two nodes may be present. Small carina posterior of azygous node extends to posterior tip, which is slightly flexed upward. Oral (upper) surface smooth to finely nodose.

Remarks.—There appears to be an almost complete gradation between this species and *P.* (*P.*) *perlobata*. Many specimens in the Utah material could be assigned to either species.

Occurrence.—Zone B (Table 1)

Subgenus *Palmatolepis* (*Manticolepis*) Müller

Type species: *P. subrecta* Miller and Youngquist, 1947

Hass (1959, p. 369, 380) considers *Manticolepis* a junior synonym of *Palmatolepis*. Although article 6 of the International Rules of Zoological Nomenclature (*in* Schenk and McMasters, 1956) points out that "from a nomenclatural standpoint . . . subgeneric and generic names . . . are of the same value," the writers can find no published account where *Manticolepis* and *Palmatolepis* are confused. *Manticolepis* has utility as an easily recognized subgeneric name, and as long as it is used as such it should be considered valid.

Species of *P.* (*Manticolepis*) are distinguished by their large broad plate and by the fact that the posterior end is bent downward. They are most characteristic of the *Manticoceras* beds in Europe and North America.

Palmatolepis (*Manticolepis*) *foliacea* (Youngquist)

(Pl. 1, fig. 13)

Palmatolepis foliaceus YOUNGQUIST, 1945, Jour. Paleontology, v. 19, p. 364, Pl. 56, figs. 11, 12
Palmatolepis foliacea MÜLLER AND MÜLLER, 1957, Jour. Paleontology, v. 31, p. 1102, Pl. 140, figs. 6–9
Palmatolepis confusionensis CHAPMAN, 1958, M.S. Thesis, Univ. Kansas, p. 32, Pl. 1, figs. 1, 2, 4, 5

Plate broad triangular to semiovate, short free blade, carina slightly curved posterior of azygous node. Posterior end strongly flexed downward. Prominent sculpture of nodes and low ridges which form pattern on margin of plate. Sigmoidal keel, broad crimp on aboral (lower) surface.

Remarks.—This species is one of the most abundant of the forms found. Hundreds of specimens were obtained, all of which have a large quantity of quartz adhering to the upper surface.

Occurrence.—Zone A (Table 1)

Palmatolepis (*Manticolepis*) *marginata* (Stauffer) var. A

(Pl. 1, figs. 9–11)

Palmatolepis marginatus STAUFFER, 1938, Jour. Paleontology, v. 12, p. 437, Pl. 53, figs. 3, 7, 8, 13, 17

Small thin warped forms with more or less rounded triangular plates and short low blade. Inner side of plate meets blade about at anterior end. Slightly curved carina consists of three to four large low nodes posterior of azygous node and does not extend to posterior end of plate which is flexed downward. Oral (upper) surface is smooth. Thin sigmoidal keel on aboral (lower) surface, secondary keel faint, crimp broad.

Remarks.—These forms are most similar to the original forms described by Stauffer (1938). They have been found only in zone A and are distinct from var. B specimens found in zone B.

Occurrence.—Zone A (Table 1)

Palmatolepis (*Manticolepis*) *marginata* (Stauffer) var. B

(Pl. 2, figs. 12, 13)

Small thin plate with triangular lobed inner side, narrow outer side parallel to carina which is straight until about four denticles anterior of azygous node where it is curved. One or two small nodes extend as small carina posterior of azygous node but meet tip in only a few specimens. Blade

is free on either side for one-third of entire length. Posterior end is straight in young forms and flexed downward in mature specimens.

Remarks.—This form is a younger evolutionary stage of *P.* (*M.*) *marginata* than is the holotype, which is most similar to var. A. Reduction in size of plate gives rise to a larger free blade and also a reduction in size and number of carina denticles. The posterior end of the plate is straight in young forms and points downward only at maturity. Complete ontogenetic series of var. A and var. B have been compared; differences are not of specific value, but they are of some stratigraphic importance.

Occurrence.—Zone B (Table 1)

Palmatolepis (*Manticolepis* ?) *quadrantinodosalobata* (Sannemann)

(Pl. 1, figs. 6–8)

Palmatolepis quadrantinodosalobata SANNEMANN 1955a, Neues Jarbuch Geologie Paläontologie Abh., Band 100, p. 328, Pl. 24, fig. 6

Broad plate with well-developed inner lateral lobe and heavy nodose sculpture on anterior part of outer side of plate. Carina is sigmoidal and in some specimens extends to posterior end which is pointed only slightly upward with a downward deflected tip. Azygous node is large and conical. Aboral (lower) surface is marked by a low keel and small crimp.

Occurrence.—Zone B (Table 1)

Palmatolepis (*Manticolepis*) *subrecta* (Miller and Youngquist)

(Pl. 1, fig. 12)

Palmatolepis subrecta MILLER AND YOUNGQUIST, 1947, Jour. Paleontology, v. 21, p. 513, Pl. 75, figs. 7–11

Palmatolepis triangularis CHAPMAN, 1958, M.S. thesis, Univ. Kansas, p. 35, Pl. 1, fig. 3

Elongate plate with short free blade, strongly sigmoidal carina; secondary carina present. Coarse nodes have uniform distribution on plate. Inner lobe is triangular. Denticles on carina are large and discrete. Posterior tip bent downward and pointed.

Occurrence.—Zone A (Table 1)

Subgenus *Palmatolepis* (*Deflectolepis*) Müller, 1956

Type species: *P. deflectens* Müller

Hass (1959, p. 369, 379) suggests that *Deflectolepis* is a junior synonym of *Panderodella*. This has been discussed under *Palmatolepis; Panderodella* is considered a *nomen dubium*.

Species of this subgenus are distinguished by the very small plate and the straight posterior end. The derivation of species through reduction in size is well documented (Müller, 1956a, p. 31–32).

Palmatolepis (*Deflectolepis*) *coronata* Müller

(Pl. 1, figs. 1–5; Pl. 2, fig. 11)

Palmatolepis (*D.*) *coronata* MÜLLER, 1956, Senck. Naturf. Gesell. Abh. 494, p. 31, Pl. 10, figs. 17, 18

Small triangular plate with large blade, free for most of its length. Posterior of azygous node the carina is faint and does not reach the straight posterior tip. Oral (upper) surface is smooth except for margin of plate which is sculptured with small ridges. Aboral (lower) surface has small crimp and a faint pit beneath azygous node.

Remarks.—There appears to be a complete gradation of forms with smooth margins to those with ridged sculpture. In Europe *coronata* has not been reported with *P.* (*D.*) *minuta*, but they occur together in the Great Basin faunas. This species may have only varietal difference from *minuta*.

Occurrence.—Zone B (Table 1)

Palmatolepis (*Deflectolepis*) *minuta* (Branson and Mehl)

(Pl. 2, fig. 1)

Palmatolepis minuta BRANSON AND MEHL, 1934, Univ. Missouri Studies, v. 8, p. 236, Pl. 18, figs. 1, 6, 7

Small triangular to bilaterally symmetrical plate, large free blade with carina extending posterior of large azygous node but not to tip which is straight. Oral (upper) surface smooth; on some specimens the margin of the plate is bent upward. Small keel on aboral (lower) surface.

Remarks.—This species occurs in zones B and C but is most abundant in zone B. Zone C specimens have margins bent upward, a feature not present on those forms in zone B.

Occurrence.—Zones B and C (Table 1)

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STRATIGRAPHY AND PALEONTOLOGY OF A LATE PLEISTOCENE BASIN, HARPER COUNTY, OKLAHOMA

By John J. Stephens

Abstract

Pleistocene (Illinoian) lake deposits are exposed in the Doby Springs area, Harper County, Oklahoma. The collapse basin in which these beds were deposited resulted from solution of salt or anhydrite in the underlying Permian strata. The rim of the basin is made up of the Permian Whitehorse group, the Pliocene Ogallala formation, and the Pleistocene (Kansan to Yarmouth) Crooked Creek formation.

In Kansan time a southeastward-flowing stream cut through the Ogallala formation into the Permian strata. This stream deposited the gravels, sands, and silts of the Crooked Creek formation in its valley. During Yarmouth time a caliche formed in the top of the formation. In Illinoian time a collapse basin formed in the Doby Springs area. Lake sediments containing the Doby Springs local fauna were deposited in this basin. Settling and collapse along the southern edge imparted the present dip to the lake beds. In Sangamon time the area was drained, and a caliche formed in the upper part of the lake sediments; later, further collapse occurred to the north and imparted the northward dip to the northern part of the initial basin deposits. The present outward radial drainage results from the collapse on the margins of the initial basin. The alluvial deposits along the modern streams are of Wisconsin and Recent age.

The Doby Springs local fauna contained in the Illinoian lake deposits consists of the remains of ostracods, mollusks, fishes, amphibians, reptiles, birds, and mammals. The seven fishes (white sucker, creek chub, flathead chub, fathead minnow, black bullhead, green sunfish, and yellow perch) indicate a cool, upstream, lake environment with smaller streams running into the lake.

Twenty-four forms of mammals from the Doby Springs local fauna are discussed, and a new species of the mammalian genus, *Peromyscus*, is described.

The following extant small mammals indicate a marsh environment with surrounding lowland meadows and some trees: arctic shrew, masked shrew, northern water shrew, shorttail shrew, muskrat, meadow vole, and jumping mouse.

The climate in the area is determined by plotting the present-day distribution of the living forms and arriving at an area of overlap. This area of overlap for the species recorded in the Doby Springs local fauna lies in southeastern North Dakota and northeastern South Dakota. The climate in Oklahoma during Illinoian time was probably similar to that in the area of overlap today. Lower summer temperatures and more effective moisture are indicated.

The Doby Springs local fauna is correlated with the Berends local fauna, Beaver County, Oklahoma, and the Butler Spring local fauna, Meade County, Kansas.

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Introduction

General Statement

Circulating ground water has removed much salt and anhydrite from the Permian strata in northwestern Oklahoma and southwestern Kansas, resulting in collapse of the overlying material. Many of the small collapse basins formed were subsequently filled with lake sediments. Many of these lake deposits contain fossils of animals that lived in and around the lake at its time of existence. This study deals with one of these small basins at Doby Springs, Harper County, Oklahoma, its deposits and their mammalian fauna.

Rocks of Permian, Pliocene, and middle to late Pleistocene age are exposed in the Doby Springs area. The lake deposits that contain the described fauna are Pleistocene (Illinoian) in age. Their glacial age is indicated by the fact that the closest modern fauna consisting of the species represented in the Doby Springs local fauna is found now 600 miles to the north in southeastern North Dakota. The presence of these forms in Oklahoma during Illinoian time indicates a cooler and moister climate than today.

Location and General Description of the Area

Doby Springs is located in Harper County in the northwestern part of Oklahoma (Fig. 1). The area studied for this report consists of 14 square miles in T. 27 N., R. 24 W., and T. 28 N., R. 24 W., located primarily in the Doby Springs township approximately 9 miles due west of Buffalo, Oklahoma, the county seat.

Doby Springs was named after C. C. Doby who homesteaded on the land at the time the Cherokee Strip was opened to settlement in 1889. C. C. Doby undoubtedly chose to homestead at this place because of the good supply of water from artesian springs. By 1920 a fair sized town (Doby Springs) was present at the area of the springs; in many places more than one family occupied every 160 acres.

Doby Springs depopulated during the years of drought in the 1930's. All that remains of the original town is the collapsed Doby barn. Two families, F. N. Miller and Mrs. L. Dees, have remained and have built homes in the area. The Doby Springs area is still of importance to the county in that all the water for the town of Buffalo now comes from the basin via a pipe line.

Physiography

The Doby Springs area lies in the Plains Border section of the Great Plains Province as described by Fenneman (1931, p. 24). He states that in the vicinity of the escarpment along the east edge of the High Plains the area is characterized by large, deep, abrupt basins owing to solution of salt, or in some places anhydrite, in the underlying rocks. Frye and Schoff (1942) discuss the formation of these basins in the western Kansas-Oklahoma area. These basins may be many miles wide and long as are the Englewood and Sitka basins to the north. The large salt flats along Cimarron River near the eastern border of the county show that a large amount of salt has been dissolved by circulating ground water and later precipitated on the surface.

Harper County is drained by Cimarron and Beaver rivers. The central part of the county, in which the Doby Springs area lies, is drained by Buffalo Creek, a tributary of Cimarron River. Doby Creek and Fish Creek both drain the Doby Springs area and flow eastward into Buffalo Creek. Beaver River drains the southwestern part of the county and is separated from the Doby Springs drainage by a high divide.

Water Supply

The town of Buffalo has drilled six wells in the Doby Springs area in the S½SW¼ sec. 10, T. 27 N., R. 24 W. These wells penetrate the Pleistocene Stump Arroyo member of the Crooked Creek formation. Water is pumped to the town through a pipe line. The last well was drilled in 1956 and unlike the five wells drilled earlier did not penetrate the Permian; it has provided the best supply of water.

The aquifer consists of sands and gravels which crop out in secs. 5 (to the west), 2, and 11 (to the east), T. 27 N., R. 24 W., and sec. 33, T. 28 N., R. 24 W., to the west. The main area of recharge seems to lie mainly in secs. 5 and 33, even though some water must come from farther west and also from the exposures in secs. 2 and 11. Both Myers (1959) and the writer failed to find any outcrops of this member farther to the east in the county other than those in secs. 2 and 11. The recharge area, thus, seems quite small and is too limited to supply a major industry in the area. In the eastern part of the county only waters rich in gypsum and salt have been encountered in drilling for oil. Therefore, any large amount of additional fresh water for use in the Buffalo area must come from Beaver and Cimarron rivers.

Field and Laboratory Work

Arthur J. Myers of the Oklahoma Geological Survey and the Department of Geology, University of Oklahoma, first noted the Pleistocene fossil-bearing beds while mapping the county. In the summer of 1955 he and Claude W. Hibbard of The University of Michigan Museum of Paleontology collected vertebrate and invertebrate fossils from these beds. The fossil-bearing beds were sampled by Hibbard and field party in the summer of 1956. In the summer of 1957 The University of Michigan field party, under the direction of Doctor Hibbard and including Robert Carroll, William Melton, Gerald Paulson, David Winsted, Michael Woodburne, and the author, collected from the deposits during most of June, July, and the first part of August. Fourteen tons of matrix was removed, taken back to Meade County, Kansas, and washed for fossils.

The author spent the summer of 1958 mapping the geology of the Doby Springs area. The

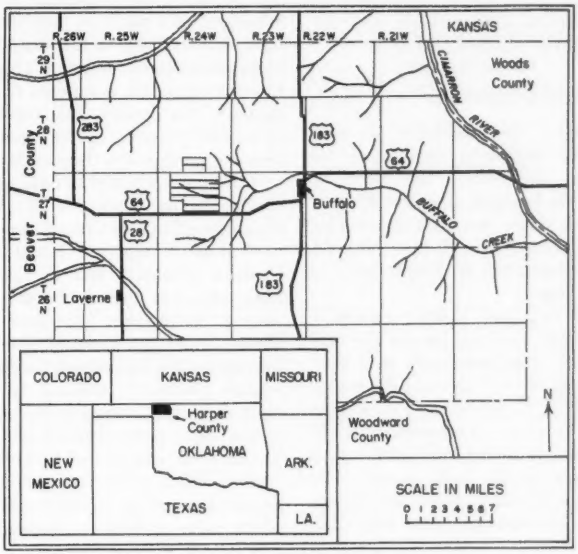


FIGURE 1.—INDEX MAP OF HARPER COUNTY, OKLAHOMA
Location of mapped area (Pl. 1) indicated by line pattern

area was visited by Dr. Donald F. Eschman and Doctor Hibbard in early September, 1958.

The laboratory and curatorial work on the fossil collection was done at the University of Michigan Museum of Paleontology. The specimens were compared with Recent specimens at The University of Michigan Museum of Zoology and with fossil material at The University of Michigan Museum of Paleontology and the American Museum of Natural History.

Acknowledgments

In the field and in the laboratory, the help of Dr. Claude W. Hibbard, chairman of the writer's doctoral committee, has been of greatest value. The original manuscript, from which this paper is taken, was critically read by Profs. William H. Burt, John A. Dorr, Donald F. Eschman, and Kenneth K. Landes of The University of Michigan.

Appreciation is expressed to the ranchers in the vicinity of Doby Springs, Harper County, Oklahoma. Mrs. L. Dees graciously permitted the author to work on her land and to remove the fossil material. F. N. Miller and J. L. Rogers gave permission to study the deposits exposed on their land and provided the author with much background on the local history. The author extends his thanks to the other citizens of Harper County who provided much assistance.

Financial support for the detailed mapping was made available in 1958 through the Horace H. Rackham School of Graduate Studies of The University of Michigan, Graduate Student Research Fund (Grant 244). The line drawings of specimens by Michael O. Woodburne (M. O. W.) were paid for by the National Science Foundation (Project G-5635). Other drawings of specimens were made by Mrs. Bonnie Hall (B. H.), Museums' artist.

The author is indebted to the American Museum of Natural History, New York, for permission to study and illustrate specimens. The writer is also grateful to Marie Hopkins of Idaho State College for measurements of jaws of the giant bison.

Stratigraphy

General Statement and Summary

Permian, Cretaceous, Pliocene, and middle to late Pleistocene deposits crop out in Harper County, Oklahoma. In the eastern two-thirds of the county Permian "red beds" are usually found at the surface. These beds belong to the El Reno and Whitehorse groups of Guadalupian age and to the Cloud Chief formation of Ochoan age (Myers, 1959). These rocks are mainly clay shales with numerous intercalated beds of dolomite, fine-grained sandstone, siltstone, and gypsum. The strata have a general southerly dip, although local collapse structures have developed as a result of underground solution. In the western part of the county Pliocene and Pleistocene deposits outcrop. The early Pliocene Laverne formation is exposed in the southwest corner of the county. The Ogallala formation of middle Pliocene age is exposed in a band trending north-south through the southwestern part of the county and in outliers on top of the divides to the northeast. Exposures of the Pleistocene Crooked Creek formation of Kansas to Yarmouth age are found overlying the Ogallala formation and the Permian Whitehorse group in the northwest portion of the county. Pleistocene (Illinoian) lake deposits overlie either the Crooked Creek or Permian formations in local collapse basins. The extreme western portion of the county is covered by dune sand of Wisconsin to Recent age, and the present-day stream valleys are filled with alluvium of Wisconsin and Recent age.

This report deals with one of the collapse basins, located in the Doby Springs area, in which rocks of Permian, Pliocene, Pleistocene, and Recent age are exposed.

Paleozoic

Permian system, Whitehorse group.—The Whitehorse group, belonging to the Guadalupian series of Permian age, is well exposed in the Doby Springs area. Clay shales and siltstones of the Whitehorse group crop out along the south side of Doby Creek in secs. 15 and 16, T. 27 N., R. 24 W. These rocks are also exposed at the surface in secs. 2 and 3, T. 27 N., R. 24 W., and 33 and 34, T. 28 N., R. 24 W. (Pl. 1). Myers (1959) measured 200 feet of exposed beds in the county that belong to this group. The exposures along the south side of Doby Creek are in a cutbank which ranges from 15 to 40 feet in height. The shales and siltstones are here capped by the Ogallala formation. The general dip along Doby Creek, resulting mainly from collapse, is to the north, measuring approximately 12°. In the other exposures the beds are nearly horizontal or dip gently in various directions (Fig. 3F). The Permian beds make up the basement rock in the area

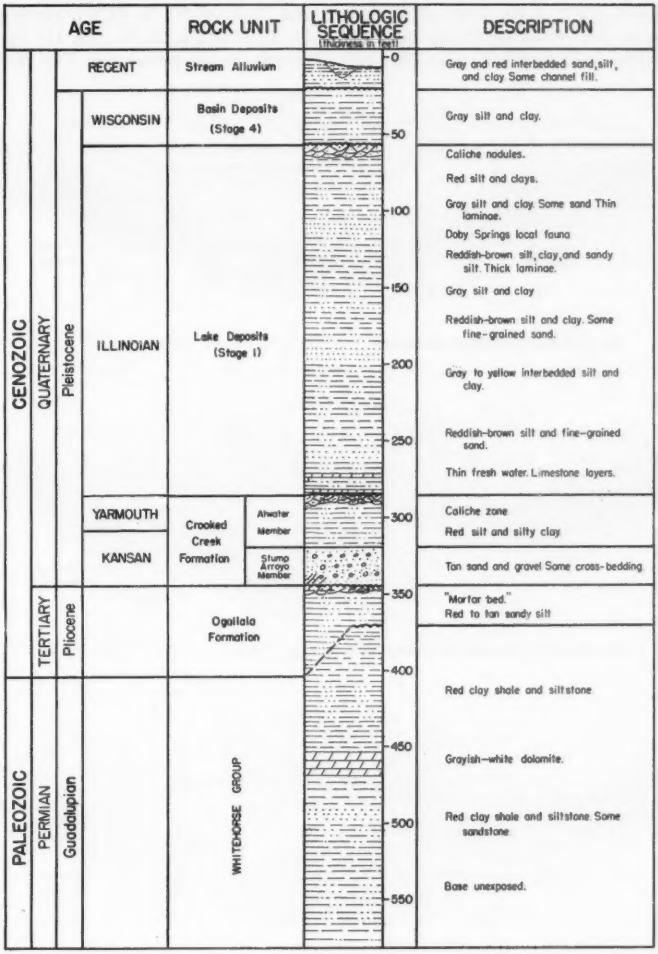


FIGURE 2.—COMPOSITE STRATIGRAPHIC COLUMN, DOBY SPRINGS AREA, HARPER COUNTY, OKLAHOMA

and form the rim rock, in part, of the Doby Springs basin. The distinct color of these beds greatly facilitated the mapping of the more recent deposits.

Cenozoic

Pliocene, Ogallala formation.—Along the south side of Doby Creek in secs. 8, 15, and 16, T. 27 N., R. 24 W., as much as 30 feet of reddish-brown to tan, calcareous, sandy silts of the Ogallala formation outcrops. In section 8 the Ogallala outcrop is represented by a "mortarlike" bed. These deposits have been mapped in part as Ogallala and in part as Crooked Creek (Myers, 1959). David B. Kitts, of the Department of Geology at the University of Oklahoma and the Oklahoma Geological Survey, who spent some time in the area, agreed with the assignment of these beds to the Ogallala formation. The Ogallala formation also outcrops approximately 250 feet above the present bed of Buffalo Creek, capping the divide between Cimarron River and Buffalo

Creek, 6 miles north of the town of Buffalo. Pliocene seeds and vertebrate fossils are found in the Ogallala rocks in other areas. No seeds were found in the beds in the Doby Springs area, but a large *Testudo* sp. toe bone and carapace fragments were recovered.

The Ogallala beds overlie Permian strata south of Doby Creek and form part of the southern rim of the Doby Springs basin (Fig. 3F).

Pleistocene, Crooked Creek formation.—Up to 29.5 feet of tannish sands and gravels of the Stump Arroyo member of the Crooked Creek formation outcrops in the E½ sec. 5, in secs. 2 and 11, T. 27 N., R. 24 W., and in part of sec 33, T. 28 N., R. 24 W. (Pl. 1). The sands and gravels are poorly sorted. The pebbles in the gravel, well rounded by stream action, average 1½ inches in diameter, with some pebbles as large as 2½ inches in diameter. In several layers of fairly well-sorted sand exposed in sec. 5, cross-bedding is preserved. The basal part of the member in the northwest corner of sec. 11 contains clay balls probably resulting from erosion of the Ogallala formation. Wherever it is directly observable the Stump Arroyo member unconformably overlies the Permian beds. This relationship is supported by water-well records.

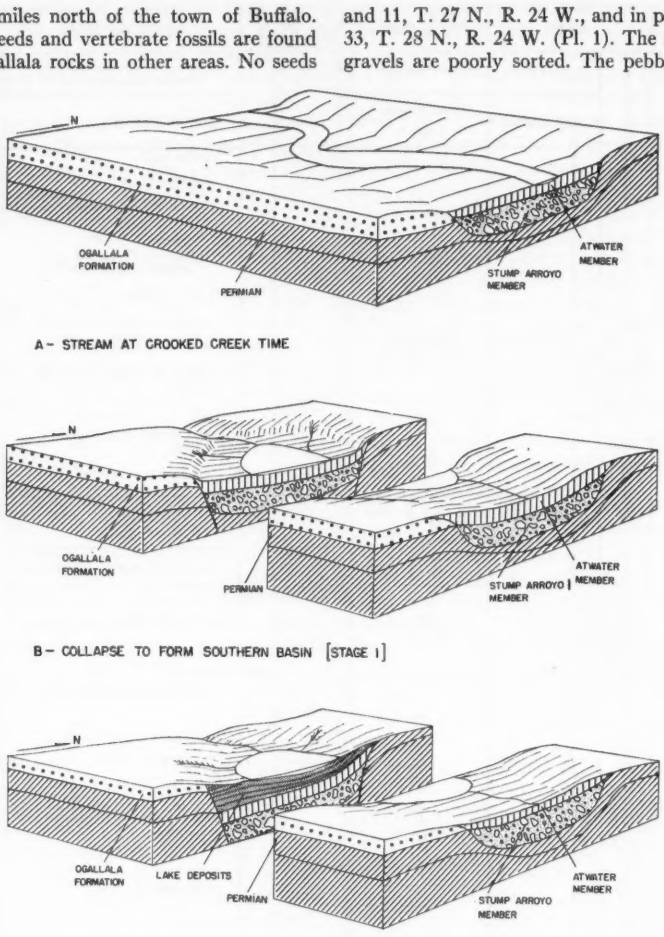
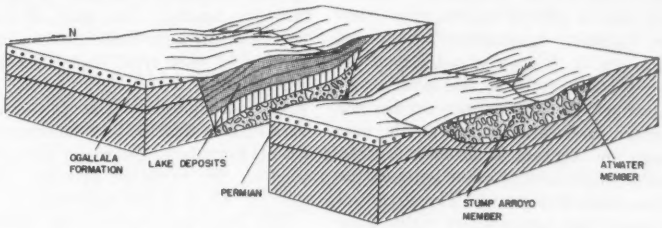
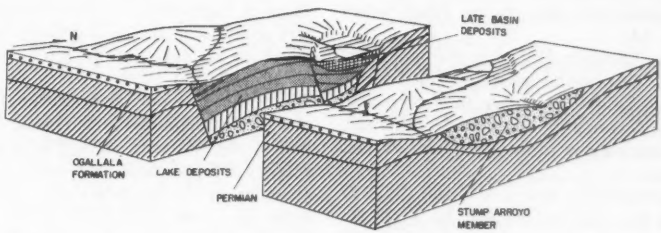


FIGURE 3.—GEOLOGIC HISTORY, DOBY SPRINGS AREA
Showing five stages of development. Area covered approximates map area (Pl. 1)

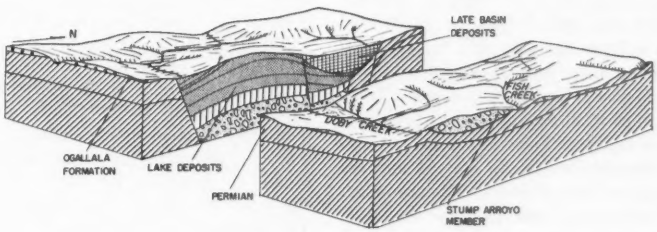
The present elevation of the Stump Arroyo member in the area indicates that a stream flowed from the west and northwest across the in a general eastward direction across Oklahoma and Kansas from approximately the same source area during Crooked Creek time,



D- FURTHER TILTING TO THE SOUTH, CALICHE FORMATION BEGINS [STAGE 3]



E- COLLAPSE AND FILL, NORTHERN BASIN. PRESENT DRAINAGE BEGINS TO FORM [STAGE 4]



F- PRESENT-DAY DRAINAGE AND STRUCTURE [STAGE 5]

FIGURE 3—D-F

Doby Springs area (Fig. 3A). This stream probably removed the Ogallala formation and cut into the Permian strata. The stream then deposited the sands and gravels in its valley. The sands and gravels are not found overlying the Ogallala beds south of Doby Creek, indicating that the area south of Doby Creek was an upland during Crooked Creek time.

The Stump Arroyo member in the Doby Springs area is probably not a continuation of deposits found in Meade County, Kansas (Hibbard, 1949a), because many streams flowed and similar deposits would be expected. Pebbles in the Stump Arroyo gravel at Doby Springs are somewhat smaller than those found in Meade County to the northwest.

Careful examination of the deposits yielded two teeth of the horse, *Plesippus simplicidens* (Cope), one an upper left premolar (U.M.M.P. no. 39479, Fig. 7G) and the other a fragmentary molar.

The Stump Arroyo sands and gravels are the aquifer for the Doby Springs area.

The Atwater member of the Crooked Creek

formation outcrops in the NW¼ sec. 5 and in the SW¼ sec. 8, T. 27 N., R. 24 W. (Pl. 1). The beds consist of reddish-brown silts and silty clays with much intermixed caliche. Twenty-six feet of silts of this member overlying the Ogallala "mortarlike" bed was measured in sec. 5. A zone of red clayey silt occurs near the top of the section here.

The Atwater silts apparently overlie the Stump Arroyo sands and gravels to the east even though the contact is nowhere visible. No noticeable unconformity exists between the silts and the sands and gravels. The stream of Crooked Creek time probably partially filled its valley with the Stump Arroyo sands and gravels and then, with a decrease in stream velocity and a concomitant decrease in particle size, deposited the silts on the broad valley floor (Fig. 3A).

A caliche has since formed in the upper part of the silts. This caliche probably formed during a warm semiarid time; it has been correlated with the Yarmouth interglacial (Hibbard, 1949a). This caliche forms the plains surface to the west.

LAKE DEPOSITS: In secs. 3, 4, 9, and 10, T. 27 N., R. 24 W., and secs. 33 and 34, T. 28 N., R. 24 W., there is an extensive accumulation of lacustrine deposits, consisting mainly of well-sorted silts, silty clays, and some sands (Fig. 2). The area of these deposits is roughly elliptical (Pl. 1). The lake beds belong to two related basins but differ in age (Fig. 3C, E). The beds of the older basin are located primarily in secs. 9 and 10 and also in the S½ secs. 3 and 4. These beds show dips resulting from postdepositional settling and collapse. The beds of the younger basin are located in secs. 33 and 34 and the northern half of secs. 3 and 4. These younger beds are horizontal and overlap the older dipping beds. It was impossible to measure stratigraphic sections in the younger beds because of vegetation, wind-blown material, and the absence of extensive dissection.

A few feet of lake deposits overlies the Stump Arroyo sands and gravels in sec. 5, T. 27 N., R. 24 W. This might be an outlier of the original lake silts or merely a small local pond deposit.

About 7 feet of horizontal beds is exposed in the NE¼ sec. 9. These deposits belong to the older basin series and consist of gray silts interbedded with layers of fresh-water limestone three-quarters of an inch to 1 inch thick. These beds probably represent early lake deposits in the center of the original collapse that were not affected by the later marginal settling and collapse (Fig. 3F).

In the beds of the older basin the layers are of limited extent and seem to grade laterally into one another. The most striking characteristic of the beds is their color. The beds, regardless of rock type, show an alternation of superimposed red and gray zones. The change in color from zone to zone is interpreted as a variation in chemical environment across an Eh-pH boundary. The writer did not determine whether the variation in color was a result solely of Eh or of pH variation.

The lake beds overlie the Stump Arroyo member as is shown by water-well records. Atwater silts are probably present between the Stump Arroyo member and the lake deposits, but it is impossible to separate them from the lake deposits on the local drilling records. The depth to aquifer (Stump Arroyo member) is usually all that local ranchers record. These figures thus give only an approximate thickness of sediment deposited in the basin. In a drilled well in the NW cor., NE¼NW½ sec. 10, the silts are 107 feet thick. The surface beds in the area dip to the northeast and belong to the upper gray zone (units 12–22, measured section below) of the older basin. The Buffalo town wells drilled in the SE¼SW¼ sec. 10 show approximately 13 feet of sediment between the surface and the aquifer. This thickness probably is made up, in part, of stream alluvium.

Because the lake sediments are of local extent they were derived probably from the Crooked Creek and Ogallala formations which outcrop around them. Small streams undoubtedly carried much sediment into the lake, but owing to the extensive later erosion around the border of the basin, no evidence of the exact location of the streams was found. The remainder of the sediment was probably carried into the lake by slope wash.

The Doby Springs local fauna is confined to one of the moderately dipping gray zones of the older basin. The fauna consists mainly of mollusks and microvertebrate material. Fragmentary elephant and horse remains were the only fossils recovered from the red zones in this basin. Mollusk shells were found in a younger calcareous silt in the NE cor., sec. 10. No fossils were found in the deposits of the northern, younger basin.

Caliche nodules are found in the top of the lake deposits and as a surface rubble both in the S½NW¼ sec. 10 and along the slope to the north in the southern part of secs. 3 and 4. The

caliche is associated only with the older lake deposits and not with the younger beds to the north. Bretz and Horberg (1949, p. 509) state:

"The profile (caliche) development is believed to be best explained by conditions of alternating saturation and desiccation in a relatively dry climate with periodic rains. Solution occurred as the soil waters moved downward during a rainy period, and precipitation took place with the capillary rise of saturated alkaline solutions as the soil zone dried out."

The presence of the caliche in the lake deposits indicates a semiarid period with good drainage after the deposition of the lake sediments and before deposition of the younger basin deposits to the north.

The age of the lake deposits of the southern basin is Illinoian as determined by stratigraphic position, faunal evidence, and correlation with other faunas and caliches. The lake deposits overlie the Crooked Creek formation of Kansan to Yarmouth age and thus must be younger than Yarmouth age.

Hibbard and Taylor (1960) describe another buried caliche from Meade County, Kansas, that is younger than the Crooked Creek caliche bed. They assign this later caliche formation to late Sangamon time. Beneath this caliche they have recovered warm- and cool-climate faunas which they correlate with Sangamon and Illinoian time, respectively. The caliche found in the Doby Springs area is believed to be the same age as the Sangamon caliche in Meade County. The presence of the Doby Springs caliche in the top of the beds and as a surface rubble shows that it could not be a result of Recent caliche formation. Recent caliche begins to form in the shape of isolated nodules around roots and pebbles beneath the surface. The absence of a massive caliche, like that in Meade County, in the Doby Springs area is possibly a result of a high water table during the beginning of caliche-forming time.

The Doby Springs local fauna collected from the lake beds, in which the caliche developed, is composed of forms that, today, live in cooler regions of the United States. Their presence in Oklahoma in past time indicates a much cooler climate than that of the present. Such a climate would have prevailed during periods of glacial advance. Because of the position of the beds and the fauna below the Sangamon caliche and above the Yarmouth deposits, their age is Illinoian.

The deposits of the younger basin to the north overlap the older lake deposits and show no development of a buried caliche. This basin probably did not form until after Sangamon time.

A composite section, measured in the NW¼-SW¼ sec. 10, T. 27 N., R. 24 W., is given below. The beds exposed in this area belong to the alternating red and gray zones of the Illinoian lake beds. The upper units were measured chiefly in a north to south traverse and the lower units in an east to west traverse.

Unit	*Description*	*Thickness in feet*
24.	Topsoil	0.5
Lake deposits		
23.	Silt and silty clay, reddish brown to tan, blocky, thick laminae, caliche nodules	3.7
22.	Silt, light gray, frosted grains, light weight, thin laminae, some bone and shell	2.0
21.	Silt and some sand, gray, no laminae, fish fossils (locality 3)	1.0
20.	Silty clay and silts, dark gray to black at top, carbonaceous at top, thick laminae with high fossil content (locality 2 at bottom, locality 4 at top)	2.0
19.	Sand, tan to brown, fine-grained, ironstone concretions, some fossils, no laminae (locality 1)	1.4
18.	Silt, light gray, thin beds, some shell	2.5
17.	Clay and silt, dark gray to brownish gray, flaky on weathered surface, highly fossiliferous (locality 5)	5.0
16.	Silty clay, light gray, thick laminae, some shell	2.5
15.	Sand, brownish, fine-grained, ironstone concretions, caliche nodules	2.6
14.	Clay, some silt, light gray, caliche nodules, thin-bedded	2.7
13.	Sand, grades upward into gray silt, tan to brown, some ironstone concretions, no bedding visible	5.9
12.	Sandy silt, light gray, thick laminae, caliche nodules	4.9
11.	Clay and silt, brownish red at bottom, red at top, alternating beds of clay and silt, beds thin toward top, thick at bottom, frosted grains in silt	35.0
10.	Clay, reddish gray, iron staining, blocky, caliche nodules	6.2
9.	Sand, reddish brown, fine-grained, well sorted, no bedding visible	7.0
8.	Silty clay, some sand at bottom, clay greenish gray on unweathered surface, sand reddish brown, iron staining, ironstone concretions,	

Unit	Description	Thickness in feet
	some indurated sand, caliche nodules	12.3
7.	Silty-clay, brownish red, blocky fracture, large bone scrap	31.6
6.	Silt, reddish brown, well sorted, calcareous cement, poor bedding, some frosted grains	28.6
5.	Silt, tan, alternating with dark-gray clay, thin laminae, ironstone concretions, caliche nodules, horse tooth	42.6
4.	Silty clay, light gray, calcite encrustations, iron staining in streaks and specks, ironstone concretions at base, thick bed, caliche nodules	6.4
3.	Silt, brownish red, some brownish clay layers, iron staining, sandy layer 1 foot thick at top, frosted grains	22.3
2.	Clay, reddish, blocky fracture, flakes on weathered surface	5.6
1.	Silt, brownish red, bedding indistinct, fragments of large bone, iron staining. Base unexposed	7.2
	Total thickness exposed	241.5

Recent, stream alluvium.—Recent stream deposits are present along Doby Creek and Fish Creek (Pl. 1). Both these streams drain the Doby Springs area and flow southeastward into an unnamed tributary of Buffalo Creek. The deposits are a result of normal stream alluviation. The top of the alluvium lies 20–30 feet below the level of the lowest Illinoian lake beds. The deposits consist of gray to reddish-brown silts and silty clays. The red color here probably resulted partly from weathering and partly from the original color of the Permian "red beds" that supplied sediment. The dissection of the lake deposits and the direction of flow of the present drainage indicates that some of the alluvium was derived from the lake sediments. Several cut and filled channels have been noted in the alluvial deposits along Doby Creek. From one of these channel fills in sec. 14, T. 27 N., R. 24 W., an isolated mammoth (elephant) tusk and tooth were collected.

The alluvial deposits show no buried caliche development and therefore were deposited after Sangamon time. As the development of the present topography and thickness of alluvial deposits probably took a reasonably long time, the writer postulates that the dissection began during late Sangamon. The deposits are probably Wisconsin and Recent in age. Fish Creek is a relatively younger stream than Doby Creek, as shown by the fact that the valley is narrow and the thickness of deposits slight.

Geologic History

During the early Cenozoic, the Permian and Mesozoic strata were subjected to erosion. There is no evidence of pre-Pliocene Cenozoic deposition in the county. The Ogallala streams flowed across the area in middle Pliocene time depositing gravels, sands, silts, and clays. Following this deposition a period of aridity occurred, and during this time a calcareous zone formed in the top of the formation. This zone, casehardened where exposed at the surface, forms the characteristic Ogallala "mortar bed." From late Pliocene to Kansan time the area was again subjected to erosion; at least there is no evidence of deposition. Probably during this time the solution of the underlying Permian anhydrite and/or salt beds began. In early Kansan time the stream that was to deposit the Crooked Creek formation flowed across the county. It probably followed a slight sag in the center of the county caused by solution and subsequent settling of the underlying Permian beds (Fig. 3A). This stream cut through the Ogallala formation and into the Permian strata in the Doby Springs area. After cutting its valley it deposited the Stump Arroyo sands and gravels and the Atwater silts in its valley. This valley fill is the most eastern exposure of the Crooked Creek formation in the county. During late Yarmouth time a caliche formed in the upper part of the Crooked Creek formation. After caliche formation, continuing solution and collapse produced a basin in the Doby Springs area (Fig. 3B). The lake sediments were deposited in this basin. The history of the development of the basin can be divided into five major stages (Fig. 3B–F).

STAGE 1: The initial collapse apparently formed a basin with its center near the NE cor., sec. 10, T. 27 N., R. 24 W. (Fig. 3B). The southern limit lay just south of the section line between secs. 10 and 15 and along and slightly north of the line between secs. 9 and 16. The northern limit lay approximately along an east-west line across the middle of secs. 3 and 4. The eastern border was along the eastern edge of sec. 10 and the western border along the western edge of sec. 9 (Pl. 1).

This basin filled with lake sediments up to a level nearly equal to that of the top of the Crooked Creek valley fill, as shown by the present height of the lake beds. The lake sediments are exposed at least 150 feet above the bed of the present Doby Creek. The fish species present in the Doby Springs local fauna (Smith, 1958) indicate an upstream pond with a small tributary outlet. This outlet probably flowed eastward into the ancestral Cimarron River following the valley formed by the stream that deposited the Crooked Creek formation (Fig. 3B–C). The deposits containing the fossils accumulated in the lake. Filling of the lake and further solution caused settling to give stage 2.

STAGE 2: Because of the weight of the sediments deposited in the initial basin and further solution, settling and collapse of the beds took place along the southern part of the basin area, and the lake deposits were tilted to the south (Fig. 3C). This tilting continued as further solution and adjustment to the weight of the sediments took place. The movement took place along the southern wall of the basin which extended from the southeast edge of sec. 10 to the SW cor., sec. 9 (Pl. 1). The southerly dip is as high as 55° at a point 250 yards northeast of the SW cor., sec. 10. After the lake beds dropped Doby Creek cut headward into the basin area from Buffalo Creek, which drained into Cimarron River. Doby Creek continued to dissect the Doby Springs area giving rise to stage 3.

STAGE 3: Doby Creek, having cut headward from Buffalo Creek, dissected the southern edge of the lake deposits and also the E½ sec. 10. This dissection lowered the water table in the lake beds, which, accompanied by a warm, semiarid climate (Sangamon time), produced a condition favorable to caliche formation. The abundant caliche nodules in the upper lake beds indicate a well-drained area. The caliche is nodular and not a massive bed as developed in the Ballard, Crooked Creek, and Kingsdown formations in Kansas. The lack of a massive bed suggests that the lowering of the water table did not take place until a large part of the caliche-forming time had passed (Fig. 3D). After the formation of the caliche, further collapse to the north took place to give rise to stage 4.

STAGE 4: Following the formation of the caliche, solution and collapse to the north of the initial lake basin occurred (Fig. 3E). This collapse caused the northward dip of the initial lake deposits. The center of this collapse was close to the NW cor., sec. 3, T. 27 N., R. 24 W. The basin formed here was filled with sediments. These sediments are horizontal and overlie the northward-dipping beds of the initial basin. The northern basin is still filling today, as is shown by the inward drainage (Pl. 1). After the final collapse and fill the present drainage system continued to form to give rise to the final stage.

STAGE 5: The final drainage pattern formed after the collapse and filling of the northern basin (Fig. 3F). Fish Creek had begun to cut headward from Buffalo Creek probably before the collapse to the north took place. After filling of the basin had begun, Fish Creek had cut through the S½ sec. 2, T. 27 N., R. 24 W., and finally captured the drainage in the E½ sec. 3. The stream in sec. 3, flowing southward into Doby Creek, previously carved out a fairly wide valley. This valley is now intersected by the narrow steep valley of Fish Creek.

The drainage pattern developed in the deposits of the initial basin is of interest in that it differs from the normal inward drainage pattern of basins. In this basin the drainage is outward from a high center. The lake deposits are at a high elevation in the center of the area, whereas older beds forming the rim of the basin are at lower elevations (Fig. 3F). This condition probably results from a combination of two factors: (1) If the dip of the beds was produced as postulated, an initial slope resulting from collapse of the outer margins would have been imparted to the area. The drainage would have been outward from the high center in the initial lake deposits. This drainage would have radiated from the center to the basin wall made up of Crooked Creek, Ogallala, and Permian rocks. Here the stream would have been deflected along the bordering rim. This situation would have produced the present drainage system. The other factor is that the Ogallala and Crooked Creek formations both offer a good supply of ground water. Two large stock ponds in sec. 8 are fed by springs in the Ogallala formation. The Stump Arroyo member is the aquifer in the area. At the time the streams were flowing along the rim, springs may have formed along the exposures of Ogallala and Crooked Creek rocks. These springs would have added water to the stream and increased the stream's erosion power. The writer offers the combination of these two factors to explain the present drainage pattern in the area of the initial basin.

The drainage of the younger, northern basin is radial inward. This is the expected drainage in a basin that is still being filled.

Fossil Localities and Methods of Collection

The Doby Springs local fauna was collected from three layers, within 10 feet of stratigraphic section. Five localities, all in the N¼SW¼ sec. 10, T. 27 N., & R. 24 W. (Pl. 1), furnished identifiable specimens. The most extensive collection was made at localities 2 and 5. Owing to the nearness of the localities stratigraphically, the writer considers that all the fossils recovered belong to the Doby Springs local fauna. The fossiliferous zone lies in the upper gray beds of the initial collapse basin.

The stratigraphic level of each locality is indicated in the measured section. The stratigraphically lowest locality is number 5 (unit 17), and the highest is number 3 (unit 21). Locality 1 is in unit 19, and localities 2 and 4 are in unit 20. Most of the fish material was recovered from locality 3 and most of the invertebrate specimens from locality 4. The other localities provided mammalian fossils, although all groups were represented in all localities.

The condition of the specimens is fair. All specimens show some mineralization. The vertebrate material is fragmentary, consisting of partial jaws, teeth, and skeletal parts. The invertebrate material consists mainly of complete shells.

In the summer of 1955, C. W. Hibbard and A. J. Myers collected specimens from the surface in the area, recovering *Ondatra*, *Equus*, *Microtus*, *Bison*, and fossil snails. On the basis of these forms the fauna was recognized to have belonged to a glacial time. In the summer of 1956, The University of Michigan field party under the direction of Doctor Hibbard, removed a ton of matrix from locality 2 and washed it. Four species of shrews and a species of squirrel were recovered from this material. In the summer of 1957 the field party removed 14 tons of matrix from the above localities. The total fossil material collected constitutes the Doby Springs local fauna.

The method of collection described by Hibbard (1949b) has been used extensively by him in the area of southwestern Kansas and northwestern Oklahoma. An excellent succession of late Cenozoic continental faunas has been obtained in this area. Forms collected by this method in the southwestern Kansas and northwestern Oklahoma area include: oögonia of *Chara*, ostracods, snail eggs, snail shells, slug skeletons, and fish, amphibian, reptile, bird, and mammal remains.

Doby Springs Local Fauna

The Doby Springs local fauna is named after Doby Springs, Harper County, Oklahoma. The fauna, composed of ostracods, mollusks, and vertebrates, was recovered from a limited area just north of Doby Creek in the SW¼ sec. 10, T. 27 N., R. 24 W., on the land of Mrs. L. Dees.

Ostracoda

The fresh-water ostracods are being studied by Richard H. Benson of The University of Kansas.

Mollusca

Mollusks are plentiful in the deposits and are presently being studied by Dwight W. Taylor of the United States Geological Survey. The presence of *Probythinella lacustris* Baker is recognized by Hibbard and Taylor (1960). This form occurs today in the cool northern lakes and streams of southern Canada, southward into Arkansas and east to the Appalachian Mountains. Its presence in the Doby Springs local fauna indicates a cool lake environment. The only other fossil record of this form from the southwestern Kansas-northwestern Oklahoma area is in the Butler Spring local fauna, Meade County, Kansas (Hibbard and Taylor, 1960).

Vertebrata

Class Pisces

The fish material collected mainly from locality 3 was studied and described by Smith (1958). He recognized the following forms:

Catostomus commersoni (Lacép de)—white sucker

Semotilus cf. *S. atromaculatus* (Mitchill)—creek chub

cf. *Hybopsis gracilis* (Richardson)—flathead chub

Pimephales promelas Rafinesque—fathead minnow

Ictalurus melas (Rafinesque)—black bullhead
Lepomis cf. *L. cyanellus* Rafinesque—green sunfish
Perca flavescens (Mitchill)—yellow perch.

Smith (Oral communication) states that the genera and species of fishes found in the Doby Springs local fauna have a present-day distribution similar to that of the genera and species recovered from the Berends local fauna (Smith, 1954). The Berends local fauna is of Illinoian age and was recovered in Beaver County, Oklahoma (Herrington and Taylor, 1958; Hibbard, 1956; Mengel, 1952; Rinker and Hibbard, 1952; Smith, 1954; 1958; Starrett, 1956; Taylor, 1954; Taylor and Hibbard, 1955). Figure 9 shows the area of overlap of the present distributions of the fishes. Smith further states (1958, p. 180):

"The Doby Springs deposits contain several species of minnows not found in the Berends site and this probably indicates that the locality was more of an upstream, pond situation with smaller streams flowing into it. The absence of drum, gar, and channel catfish would seem to bear this out."

Class Amphibia

Remains of frog and salamander recovered from the deposits are being studied by J. A. Tihen of the University of Illinois.

Class Reptilia

Numerous snake vertebrae and several snake jaws present in the fauna are being studied by Bayard H. Brattstrom of Adelphi College. Turtle fragments were recovered but cannot be identified because they are incomplete.

Class Aves

A few bird bones from the fauna are being studied by Pierce Brodkorb of the University of Florida.

Class Mammalia

Twenty-four members of the mammalian fauna were identified. These are listed below and discussed in the following section.

Insectivora
Sorex arcticus Kerr—arctic shrew
Sorex cinereus Kerr—masked shrew
Sorex palustris Richardson—northern water shrew
Blarina brevicauda (Say)—shorttail shrew

Rodentia
Citellus cf. *C. richardsonii* (Sabine)—Richardson ground squirrel
Citellus tridecemlineatus (Mitchill)—thirteen-lined ground squirrel
Thomomys sp.—western pocket gopher
Geomys bursarius (Shaw)—plains pocket gopher
Castorid, genus and sp. indet.—large beaver
* *Peromyscus* cf. *P. cochrani* Hibbard
* *Peromyscus oklahomensis* sp. nov.
Onychomys cf. *O. leucogaster* (Wied-Neuwied)—northern grasshopper mouse
Ondatra zibethicus (Linnaeus)—muskrat
Microtus pennsylvanicus (Ord)—meadow vole
Zapus sp.—jumping mouse

Carnivora
A large form, fam. indet.
A small form, fam. indet.

Proboscidea
† *Mammuthus* sp.—mammoth

Lagomorpha
Lepus sp.—hare

Artiodactyla
† *Camelops* sp.—camel
Antilocaprid—pronghorn
* *Bison* cf. *B. latifrons* (Harlan)—giant bison

Perissodactyla
* *Equus* cf. *E. niobrarensis* Hay
* *Equus* sp.—horse

† Extinct genus and species
* Extinct species

Systematic Description of Fossil Mammalia

A systematic description is given below for each member of the mammalian fauna. Catalog numbers, if not otherwise identified, refer to specimens in the collections of The University of Michigan Museum of Paleontology (U.M.M.P.). Those specimens in the collections of the American Museum of Natural History are designated by A. M. N. H.

The classification and nomenclature used for Recent species is that of Miller and Kellogg (1955).

Class MAMMALIA

Order INSECTIVORA

Family Soricidae

Sorex arcticus Kerr

(Fig. 4F–H)

Geologic distribution.—Illinoian to Recent. This is the first fossil record of the arctic shrew recorded.

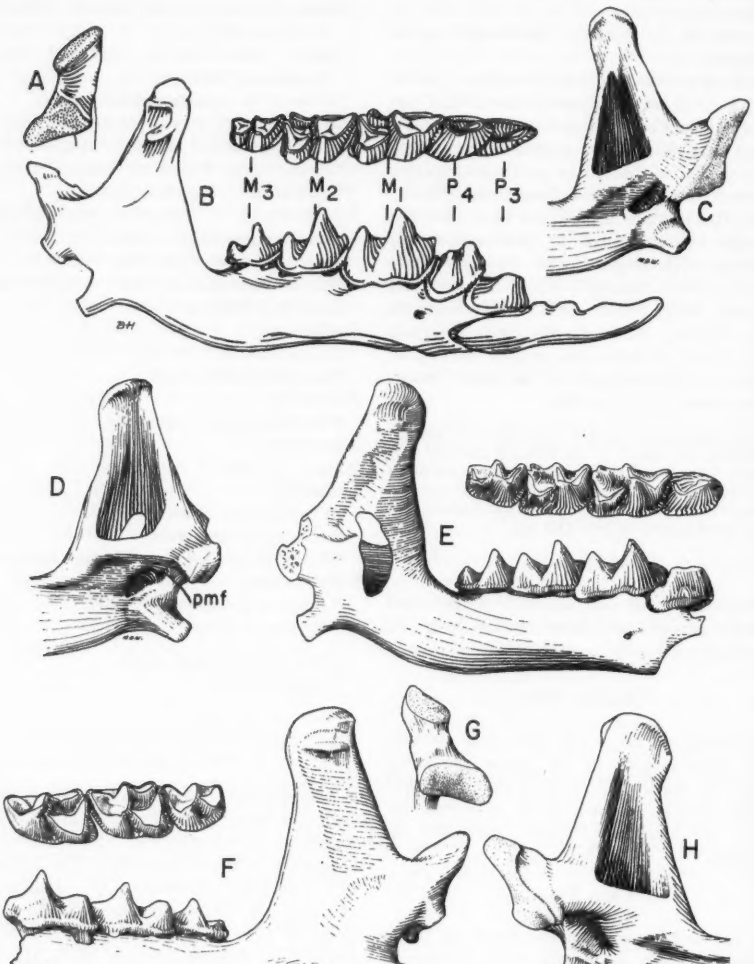


FIGURE 4.—DRAWINGS OF *SOREX*

A–C, *Sorex palustris:* U.M.M.P. 34748, right lower jaw, I–M_3; A, posterior view of condyle; B, labial and occlusal views; C, lingual view of posterior part of jaw

D–E, *Sorex lacustris:* U.M.M.P. 38361, right lower jaw, P_4–M_3; D, lingual view of posterior part of jaw; E, labial and occlusal views

F–H, *Sorex arcticus:* U.M.M.P. 38613, left lower jaw, M_1–M_3; F, labial and occlusal views; G, posterior view of condyle; H, lingual view of posterior part of jaw

pmf—post-mandibular foramen

A–E, ×8; F–H, ×10. Drawings by Mrs. Bonnie Hall (B. H.) and Michael O. Woodburne (M. O. W.)

Habitat.—The arctic shrew lives in cold damp places, under leaves along forested valleys, marshes, or lake shores (Bailey, 1926).

Material.—Doby Springs local fauna, nos. 38601, 38607–38608, 38613–38615, 21 partial lower jaws; no. 38616, two partial maxillaries

Remarks.—The fossil specimens are of a size equal to that of *Sorex arcticus arcticus* Kerr. The lower jaws compare closely with *S. arcticus* in possessing the following characters: a well-developed postmandibular foramen; indentation for attachment of masseter muscle parallel to the posterior edge of the coronoid process (Gaughran, 1954); articular surfaces similar; molars with a well-developed anterior cingulum. In some of the lower molars a ridge was present, extending from the metaconid to the entoconid (Fig. 4F). Examination of a series of Recent specimens of *S. arcticus* in the collections of the Museum of Zoology of The University of Michigan showed that 5 out of 70 possessed this ridge and that it was merely an old-age character resulting from the wearing away of the hypoconid. The length of the lower molar tooth row (M_1–M_3) is 3.5 mm in specimen no. 38613 (Fig. 4F).

In one of the maxillaries with only P^4 (no. 38616), the parastyle is separated from the metacone as in the Recent species.

There are 11 lower right jaws and 10 left lower jaws indicating at least 11 individuals.

Sorex cinereus Kerr

(Fig. 5E–F)

Geologic distribution.—Kansan to Recent. This species was reported from the Kansan Cudahy fauna by Hibbard (1944).

Habitat.—The masked shrew occupies practically every meadow, brush patch, and grove and is found under rich vegetation almost anywhere on the North American prairies north of northern Nebraska.

Material.—Doby Springs local fauna, nos. 34756, 38602–38603, 38611–38612, 38618, 11 partial lower jaws; no. 38618, 1 partial maxillary

Remarks.—The length of the lower molar tooth row (M_1–M_3) in no. 34756 (Fig. 5E) is 3.0 mm. This species does not have a postmandibular foramen in the lower jaw. The size of the jaw and the characters of the teeth are like those of the Recent *Sorex cinereus cinereus* Kerr.

The minimal number of individuals present as determined by counting left lower jaws is 8. There are 3 right lower jaws.

Sorex (*Neosorex*) *palustris* Richardson

(Fig. 4A–C)

Geologic distribution.—Illinoian to Recent. This is the first Illinoian record of this species.

Habitat.—The northern water shrew is an expert swimmer and apparently spends much time and obtains much of its food in water. It inhabits the edges of cool lakes and marshes of the northern part of the United States, north of central Minnesota and Wisconsin.

Material.—Doby Springs local fauna, nos. 34747–34748, 38599–38600, 38604–38605, 38619–38620, 18 lower jaws; nos. 34753–34754, 38606, 38609–38610, 38621, 6 partial maxillaries

Remarks.—The specimens represent a shrew the size of the Recent *Sorex* (*Neosorex*) *palustris palustris* Richardson and, from the characters of the jaws and teeth present, cannot be distinguished from that subspecies. The fossil possesses a third upper unicuspid smaller than the fourth and does not have a postmandibular foramen (Fig. 4C). The length of the lower molar tooth row (M_1–M_3) is 3.9 mm in no. 34748 (Fig. 4A), 3.8 mm in no. 34734, and 4.0 mm in no. 38599, all right lower jaws. At least 10 individuals are represented.

Hibbard (1944, p. 721) described a new species of shrew, *Sorex* (*Neosorex*) *lacustris*, from the Cudahy fauna, Crooked Creek formation, Meade County, Kansas. Additional material collected in the summer of 1958 by The University of Michigan field party from the Crooked Creek formation shows that Hibbard's form does not belong in the subgenus *Neosorex*. *S. lacustris* has a well-developed postmandibular foramen (Fig. 4D–E, no. 38361), not unlike that of *S. arcticus*. It most probably belongs within the line that gave rise to *S. arcticus*.

Blarina brevicauda (Say)

(Fig. 5C)

Geologic distribution.—Illinoian to Recent. This species is also found in the Berends local fauna, Beaver County, Oklahoma, of Illinoian age (Starrett, 1956).

Habitat.—The shorttail shrew lives in moist places in mellow soil under fallen grass and leaves. It prefers cool moist areas around woods, streams, and lakes.

Material.—Doby Springs local fauna, no. 34752, a partial left maxillary

Remarks.—A single maxillary with P^3 and P^4 of this form was found. Extensive washing an surface collecting during the summers of 1957 and 1958 failed to duplicate this form in the fauna.

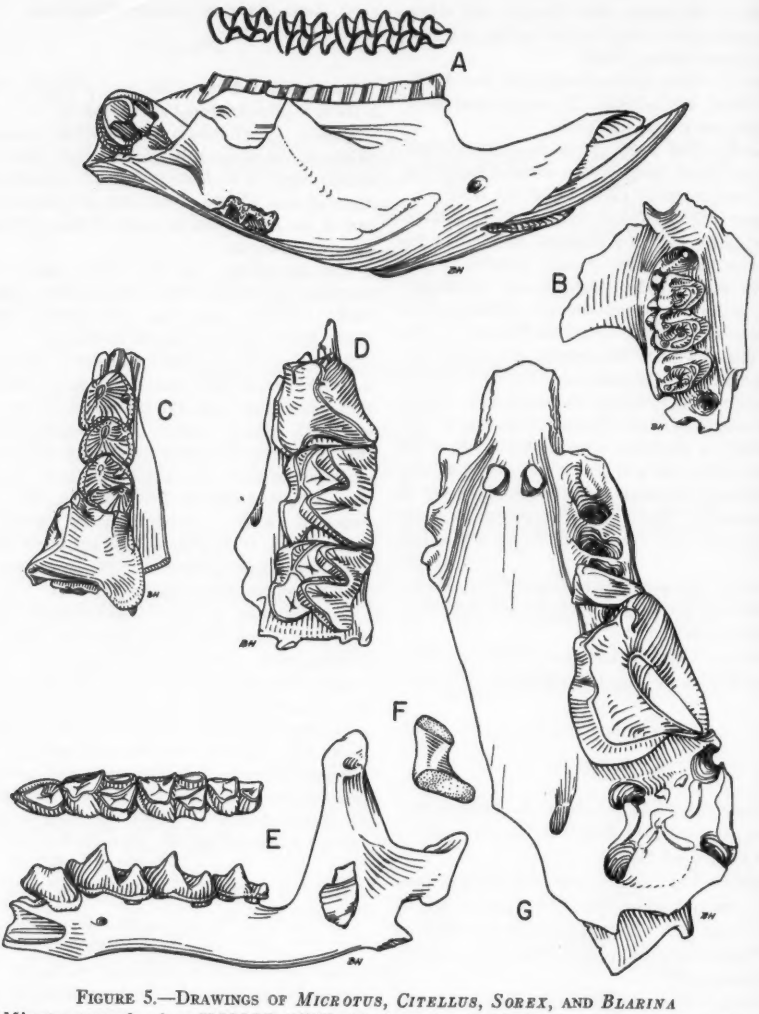


FIGURE 5.—DRAWINGS OF *MICROTUS*, *CITELLUS*, *SOREX*, AND *BLARINA*

A, *Microtus pennsylvanicus:* U.M.M.P. 34755, right lower jaw, I, M_1–M_3: lateral and occlusal views

B, *Citellus tridecemlineatus:* U.M.M.P. 34751, partial right maxillary, P^4–M^1; occlusal view

C–D, *Sorex palustris:* C, U.M.M.P. 34754, partial right maxillary, C, P^1–P^4; occlusal view; D, U.M.M.P. 34753, partial left maxillary, P^4–M^2, occlusal view

E–F, *Sorex cinereus:* U.M.M.P. 34756, left lower jaw, P_4–M_3; E, labial and occlusal views; F, posterior view of condyle

G, *Blarina brevicauda:* U.M.M.P. 34752, partial left maxillary, P^2, P^4; occlusal view

A, ×6; B, ×4; C–G, ×10. All drawings by Mrs. Bonnie Hall (B. H.)

The specimen shows no differences from the maxillaries of Recent specimens of *Blarina brevicauda* (Say) except in size. The fossil is somewhat larger than the largest *Blarina* in the collections of the Museum of Zoology of The University of Michigan. In the largest specimens in these collections (U.M.M.Z. no. 107570) the anteroposterior length of P^4 is 2.0 mm, whereas in the fossil (no. 34752, Fig. 5C) this measurement is 2.4 mm. The palate also appears broader in the fossil than in the

Recent form. The current assignment of forms to this species, regardless of size, by other workers is followed here.

More specimens are required before this form can be assigned to a subspecies of *B. brevicauda*.

Order RODENTIA

Family SCIURIDAE

Citellus cf. *C. richardsonii* (Sabine)

Geologic distribution.—Illinoian to Recent

Habitat.—Howell (1938) states, "Richardson's ground squirrel . . . lives on open prairies and on flats along the shores of lakes and rivers, seeming to prefer the sandy or gravelly ridges."

Material.—Doby Springs local fauna, no. 38648, 17 isolated teeth

Remarks.—Seventeen isolated teeth of a ground squirrel the size of *Citellus richardsonii elegans* (Kennicott) were collected. Seven were upper molariform teeth, 9 were lower molariform teeth, and 1 was a deciduous upper premolar. The teeth compare favorably in pattern and size with the teeth of the Recent form.

Citellus tridecemlineatus (Mitchill)

(Fig. 5B)

Geologic distribution.—Illinoian to Recent. This is the earliest known occurrence of this form.

Habitat.—The thirteen-lined ground squirrel lives mainly on dry grassy praries and secondarily in sandy river bottoms and the brushy borders of timber tracts (Howell, 1938, p. 18).

Material.—Doby Springs local fauna, no. 34751, a right maxillary, and no. 38645, 9 isolated teeth

Remarks.—The maxillary is the size of a maxillary belonging to *Citellus tridecemlineatus pallidus* (Allen). The antero-posterior length of the maxillary measured from the anterior edge of the alveolus of the P^3 to the posterior edge of the alveolus of the M^3 is 7.0 mm. The slight excavation anterior to P^3 is developed to the same extent as in the Recent form (Fig. 5B).

Family GEOMYIDAE

Thomomys sp.

(Fig. 7C)

Geologic distribution.—This genus is known from the upper Pliocene to the Recent in North America.

Habitat.—The western pocket gopher spends most of its life underground. It avoids wet places and lives in the high dry praries (Bailey, 1926).

Material.—Doby Springs local fauna, nos. 38576–38577, 34 isolated teeth

Remarks.—The teeth are from a pocket gopher the size of *Thomomys talpoides quadratus* Merriam. *Thomomys* can be differentiated from all other genera of the Geomyidae by the presence of an enamel plate on the posterior loop of P^4 (Fig. 7C). The upper premolars can easily be distinguished from the lower premolars by the nature of wear. In the upper tooth the anterior loop is more worn than the posterior, whereas in the lower the reverse is true.

Two of the lower molars, probably both M_1, have a third enamel plate running down the labial side of the tooth. This character is present in isolated teeth of *T. gidleyi* Wilson, from the Hagerman local fauna, and also in young individuals of the Recent genus. The molar teeth can be differentiated from those of *Geomys* because they are constricted labially in the upper molars and lingually in the lower molars. *Geomys* molars are more nearly ovate.

At least six individuals are represented in the fauna. The writer is unable to assign these teeth to a species until more nearly complete material is recovered.

Geomys bursarius (Shaw)

(Fig. 7D, F)

Geologic distribution.—Illinoian to Recent

Habitat.—The plains pocket gopher is found today in areas of extensive soil development. It prefers low sheltered areas in valleys rather than open dry upland.

Material.—Doby Springs local fauna, no. 38574, left lower jaw; no. 38575, 15 isolated teeth

Remarks.—The lower jaw is complete with the exception of the ascending ramus and a broken incisor. The anteroposterior length of the tooth row (P_4–M_3) in the lower jaw (no. 38574, Fig. 7F) is 6.5 mm. This is well inside the size range of the Recent species. The jaw is indistinguishable from the jaw of the Recent *Geomys bursarius*. Figure 7D shows a P^4 (no. 38575 a) occlusal view. This figure may be compared with Figure 7C of *Thomomys* P^4 (no. 38577) to note the difference in enamel pattern.

Family CASTORIDAE

Material.—Doby Springs local fauna, no. 38785, incisor and cheek tooth fragments; no. 38786, proximal end of right radius.

Remarks.—The fragment of enamel from an incisor and the proximal end of a right radius are from an animal the size of one of the giant beavers. The radius (no. 38786) measures 24.5 mm across the articular surface.

It is impossible to assign the cheektooth fragments to a genus because they are incomplete and because they come from an immature animal.

Family CRICETIDAE

Peromyscus cf. *P. cochrani* Hibbard

Peromyscus cochrani HIBBARD, 1955a, Univ. Mich. Mus. Paleont. Contr., v. 12, p. 210

Geologic distribution.—Illinoian and Sangamon. This species is known from the Jinglebob local fauna, Meade County, Kansas, of Sangamon age.

Habitat.—The recovery of the Jinglebob local fauna from a stream deposit (Hibbard, 1955a) indicates that this form probably lived along the moist stream banks.

Material.—Doby Springs local fauna, no. 38571, right M^1 and left M_1.

Remarks—The two teeth compare in size with teeth of a mouse the size of *Peromyscus leucopus noveboracensis* (Fischer). The tooth pattern is similar to that of *P. cochrani* Hibbard.

The anteroconid of M_1 has a median groove. The absence of this groove was used by Hibbard (1955a) as one of the characters of *P. cochrani*. The presence or absence of a slight groove in the teeth of young individuals was not established by Hibbard because of the lack of specimens in early stages of wear. Possibly a shallow groove might be present in young individuals. In the referred specimen the anterior groove is developed to a slight degree and only reflected on the external face of the enamel. There is a weakly developed mesostylid and ectostylid. The internal and external re-entrant valleys between the cusps are as wide as those in *P. cochrani*. The M^1 compares with that of *P. cochrani* in having an anterior median fold on the anterocone in youth. Both are of the same size, and both have slightly developed mesolophs.

The referred specimen is distinct from *Peromyscus berendensis* Starrett in being larger and much broader. On the M_1 of *P. berendensis* (no. 31789) there is a mesolophid that ends lingually in an enlargement that might be considered a rudimentary mesostylid. The anteroconid possesses a well-developed labial and median fold which is absent in the Doby Springs specimen and in *P. cochrani*.

Peromyscus oklahomensis sp. nov.

(Fig. 7B)

Holotype.—No. 38672, University of Michigan Museum of Paleontology, a right M_2

Horizon and type locality.—Late Pleistocene (Illinoian), Doby Springs lake beds, Doby Springs local fauna, locality no. 2, NE¼SW¼ sec. 10, T. 27 N., R. 24 W., Mrs. L. Dees Ranch, Harper County, Oklahoma

Diagnosis.—A mouse tooth the size of M_2 of *Peromyscus floridanus* Bangs, but with extremely broad re-entrant valleys between the cusps. The mesostylid is well developed, but there is no evidence of a mesolophid. The ectostylid is very rudimentary.

Description of holotype.—Specimen no. 38572 (Fig. 7B) is the right M_2 of an adult animal. The anteroposterior length is 1.68 mm, and the greatest width is 1.30 mm. A strong mesostylid is developed between the metaconid and the entoconid. A rudimentary ectostylid is present between the protoconid and hypoconid.

Discussion.—The size of this tooth compares favorably with that of the M_2 of *Peromyscus floridanus*. The median and primary folds are much wider than in *P. floridanus*. In *P. oklahomensis* the angle of the median fold is approximately 75°, whereas in *P. floridanus* and other *Peromyscus* the average angle is 50°, depending on the development of the ectostylid.

P. oklahomensis is distinguished from all other *Peromyscus* by its large size and the wide angle between the cusps. Comparison with Recent specimens of *P. floridanus* shows that the fossil form had re-entrant valleys much wider even than specimens of *P. floridanus* without an ectostylid.

The tooth of the described form is larger than the tooth of *P. californicus* (Gambel), which is one of the largest of the white-footed mice.

Possibly this mouse is related to *P. floridanus* which may have had a much greater range during the Pleistocene than it now has. Until better material is recovered it is impossible to relate this form to any other living or fossil *Peromyscus*.

Dr. Emmet T. Hooper of The University of Michigan Museum of Zoology has examined the described specimen and agrees that, to his knowledge, it is distinct from any other species of *Peromyscus*.

Onychomys cf. *O. leucogaster* (Wied-Neuwied)

(Fig. 7A)

Geologic distribution.—Illinoian to Recent

Habitat.—The northern grasshopper mouse is a prairie animal living either in forest or dense thickets

but mainly scattered over the open country in bare and exposed situations as well as under cover of grass, cacti, weeds, and low scattered brush (Bailey, 1926).

Material.—Doby Springs local fauna, no. 38570, a left M_1

Remarks.—The greatest anteroposterior length of the M_1 is 1.9 mm, and the greatest width is 1.3 mm. The tooth is from an adult animal the size of *Onychomys leucogaster arcticeps* (Rhoads). It is impossible to determine whether this tooth belongs to *O. jinglebobensis* Hibbard (1955a) which is distinguished on characters present in the jaw and the M_2. The absence of a cingulum surrounding the tooth distinguishes the tooth from one of *O. fossilis* Hibbard (1941). The tooth fits into the size ranges of *O. jinglebobensis* and *O. leucogaster*. Better material is necessary to assign this form to a given species.

Ondatra zibethicus (Linnaeus)

(Fig. 6E–J)

Ondatra zibethicus (Linnaeus), 1766, *Systema naturae*, ed. 12, v. 1, p. 79
Fiber nebracensis Hollister, 1911, N. Am. Fauna, no. 32, p. 32
Ondatra triradicatus Starrett, 1956, Jour. Paleont., v. 30 p. 1188

Geologic distribution.—Illinoian to Recent. This form is found in the Berends local fauna (Illinoian) and the Jinglebob local fauna (Sangamon) as well as the Recent.

Habitat.—Hollister (1911) states that the muskrat is a builder of marsh houses and spends most of its life in the water. It lives in lakes and along the banks of streams, and in some cases instead of building a house it burrows from under water level into the bank to above high-water mark. Here it excavates a dry den.

Material.—Doby Springs local fauna, no. 38648, left lower jaw; no. 38646, parts of lower jaws and partial maxillary associated; no. 38647, a partial maxillary; no. 33119, 66 isolated teeth.

Remarks.—A muskrat the size of *Ondatra zibethicus cinnamominus* Hollister. In the upper and lower teeth of the Doby Springs specimens the development of dentine tracts on the labial and lingual sides of the teeth and the development of interstitial cement in the re-entrant angles is the same as in the Recent species. Cement is not deposited on the teeth of younger individuals but is present at the time of root formation (Fig. 6E–J). The dentine tracts are better developed on the labial side of M_1 than on the lingual side (Fig. 6E–J). Of the eight M^1's collected five have three roots and three have two roots each. In the specimen (no. 38646) with partial lower jaw and partial maxillary associated, the following occlusal lengths were measured: M_1, 7.0 mm; M_2, 3.5 mm; M^1, 4.8 mm; and M^2, 3.7 mm. The average occlusal length of 11 fossil M_1's is 6.5 mm. The size range is from 5.6 mm to 7.3 mm, depending mainly on age The largest tooth is from a 3-year-old, or older, individual.

Hollister (1911) in his revision of the muskrats described two new species, *Ondatra nebracensis* and *O. oregonus*. Comparison of the type skull of *O. nebracensis* (A.M.N.H. no. 2702) and the referred specimens (A.M.N.H. nos. 2703–2710) with a series of Recent muskrat skulls has shown that the characters on which he separated this form from the Recent *Ondatra* are merely characters of old age. The short rostrum (A.M.N.H. no. 2702) fits well into the range of variability in the Recent form as does the height of the interorbital crest (mostly plaster reconstruction) and the small flattened parietals and interparietals. The auditory bulla is small but seems to have been elevated in restoration from a natural position and fits into the extreme minimal size range of Recent *Ondatra*. The bulla appears to be much more like a bulla from a young animal than the skull or teeth would indicate. Possibly this bulla was collected close to the type but belonged to a younger animal and in reconstruction was placed with the old skull. All the other specimens (lower jaws and partial maxillaries) are as measured by Hollister and cannot be distinguished from the Recent species.

In 1956 Starrett described a new muskrat, *Ondatra triradicatus*, from the Illinoian, Berends local fauna. The basis of description of the new form was on the presence of three roots in the M^1. In all other characters it does not differ from the Recent *O. zibethicus*. The writer was fortunate to obtain a series of 654 M^1's of known age. Most of the teeth (364) are from 1-year-old muskrats. The rest range in age up to 3 years. In the teeth of 1-year-old individuals, 85.5 per cent had two roots and 14.5 per cent had three. In the teeth from 3-year-old animals, 43.8 per cent had two roots and 56.2 per cent had three. The type of *O. triradicatus* (no. 31781) is an M^1 from about a 2- to 3-year-old individual. Further collecting should recover M^1's having two roots as well as three. On the basis of this study it seems reasonable to consider *O. triradicatus* Starrett as synonymous with *O. zibethicus*.

Better material was recovered in the summer of 1958 of *Ondatra kansasensis* Hibbard, from the Cudahy fauna, of southwestern Kansas (Hibbard, 1944). A comparison of this material (nos. 38334–38337, 2 lower jaws and 21 isolated teeth) was made

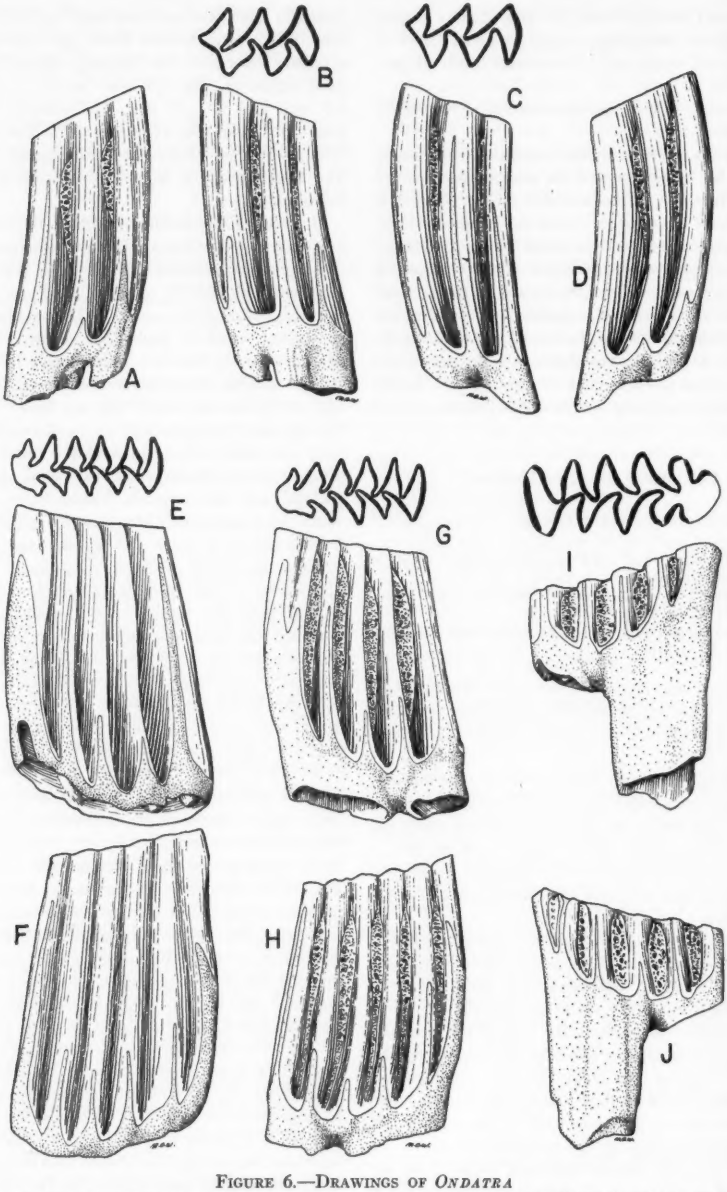


FIGURE 6.—DRAWINGS OF *ONDATRA*

A–D, *Ondatra annectens:* A–B, U.M.M.P. 38337, RM^1; A, labial view; B, lingual and occlusal views; C–D, A.M.N.H. 12424, RM^1; C, lingual and occlusal views; D, labial view

E–J, *Ondatra zibethicus:* E–F, U.M.M.P. 33119a, immature LM_1; E, labial and occlusal views; F, lingual view; G–H, U.M.M.P. 33119b, 1-year-old LM_1; G, labial and occlusal views; H, lingual view; I–J, U.M.M.P. 33119c, 3-year old RM_1; I, labial and occlusal views; J, lingual view

A–D, ×5; E–J, ×4. All drawings by Michael O. Woodburne (M. O. W.)

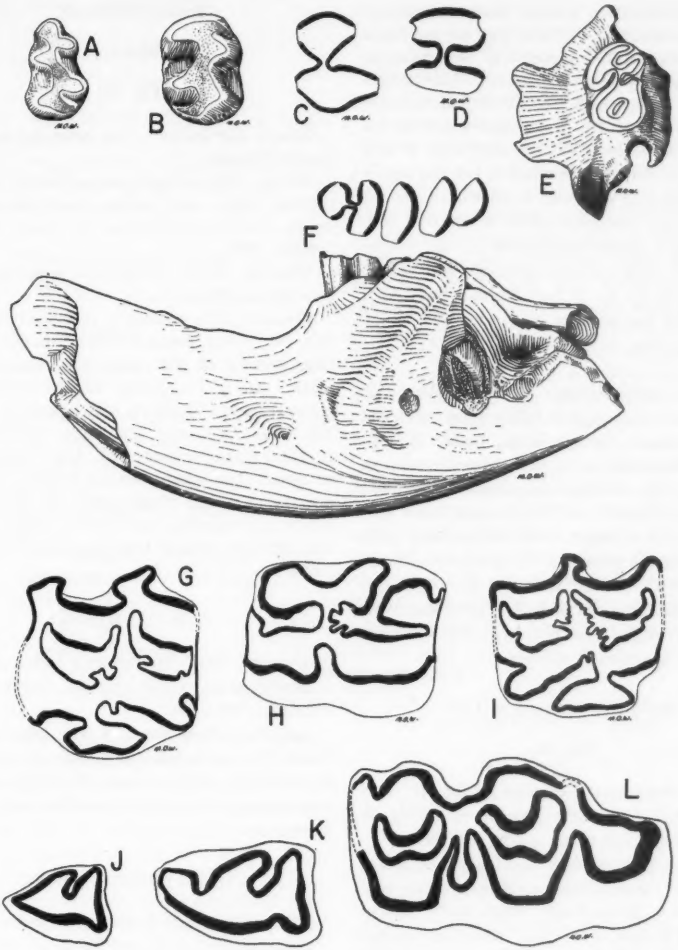


FIGURE 7.—DRAWINGS OF *ONYCHOMYS*, *PEROMYSCUS*, *THOMOMYS*

A, *Onychomys* cf. *leucogaster:* U.M.M.P. 38570, LM_1; occlusal view; B, *Peromyscus oklahomensis:* U.M.M.P. 38572, RM_2; occlusal view; C, *Thomomys* sp.: U.M.M.P. 38577, M^1; occlusal view; D, *Geomys bursarius:* U.M.M.P. 38575a, M^1; occlusal view; E, *Zapus* sp.: U.M.M.P. 38573, partial left maxillary, M^1; occlusal view; F, *Geomys bursarius:* U.M.M.P. 38574, left lower jaw, I, P_4–M_3; labial and occlusal views; G, *Plesippus simplicidens:* U.M.M.P. 39479, upper left premolar; occlusal view; H–I, *Equus* cf. *niobrarensis:* H, U.M.M.P. 38791, LP_3; occlusal view; I, U.M.M.P. 38790, RM^2 (sectioned); section pattern; J–L, *Bison* cf. *latifrons:* U.M.M.P. 38796; J, LP_2; K, LP_3; L, LM_3; all occlusal views

A, ×8; B, ×10; C–D, ×8; E, ×10; F, ×4; G–L, ×1. All drawings by Michael O. Woodburne (M.O.W.)

with *Ondatra annectens* (Brown), a left lower jaw and right M^1 (A.M.N.H. no. 12424) from the Conard Fissure, Arkansas (Brown, 1908). This comparison showed that *O. kansasensis* is the same as *O. annectens*. The anteroposterior length of M_1 of *O. annectens* is 5.7 mm, and the average of the same measurement in five specimens of *O. kansasensis* is 5.6 mm with the range from 5.4 mm to 5.7 mm. The dentine tracts are similar in both forms as is the development of the interstitial cement in the labial and lingual re-entrant angles. The depth of re-entrant angles of the anterior loop

of M_1 is a variable in muskrat lower first molars, and this variability is well shown by the specimens of *O. kansasensis*. The presence of open triangles in teeth has been shown (Galbreath, 1946) to be a variable in the muskrat. The M^1 of both compare favorably (Fig. 6A-D). The posterior end of the jaws of *O. kansasensis* was not recovered, so it is impossible to determine whether it had the square inferior notch of *O. annectens*. In some of the Recent individuals the process angularis is deflected labially, giving a square appearance to the inferior notch. The Conard Fissure probably had formed and begun to fill by at least Kansan time, as is indicated by the presence of *O. annectens* in the Kansan Cudahy fauna and the Conard Fissure deposits.

Hibbard (1959) discusses the relationship of *O. kansasensis* to *O. idahoensis* Wilson from the Grand View local fauna, Jackass Butte, Idaho. He considers *O. kansasensis* to be more advanced than *O. idahoensis*. The author examined *O. oregonus* Hollister (A.M.N.H. no. 8594, a right lower jaw) from Silver Lake, Oregon, which is larger and more advanced than *O. annectens*. The symmetrical shape of the seventh triangle of M_1 of *O. oregonus* is similar to the fifth of *O. idahoensis* (Hibbard, 1959, Fig. 7C). Possibly these two forms belong in one line and *O. annectens* in another.

Microtus pennsylvanicus (Ord)

(Fig. 5A)

Geologic distribution.—Illinoian to Recent

Habitat.—The meadow vole lives mainly in meadows, fields, and grassy places near water.

Material.—Doby Springs local fauna, nos. 38368, 38374–38375, 38380–38382, seven maxillaries; nos. 33120, 34755, 38364–38367, 38369–38373, 38376–38379, thirty-eight lower jaws; numerous isolated teeth.

Remarks.—There are 114 right lower first molars and 107 left; at least 114 individual meadow voles are represented. The fossil forms are indistinguishable from the Recent species. The anteroposterior length, M_1–M_3 (no. 34755, Fig. 5A), is 6.0 mm. The isolated M_1's include 125 with 5 alternating closed triangles and 58 with 6 closed triangles. *Microtus pennsylvanicus* does not live in the Doby Springs area today but occurs only as far south a northern Nebraska and on the east coast into South Carolina. It is found as far north as the northern Hudson Bay region.

Family ZAPODIDAE

Zapus sp.

(Fig. 7E)

Geologic distribution.—This genus is known from the late Pliocene.

Habitat.—The jumping mouse lives in meadows, shrubby fields, and thickets along the edges of woods. It shows preference for moist locations (Preble, 1899).

Material.—Doby Springs local fauna, no. 38574, a partial maxillary

Remarks.—The specimen recovered was of a very old animal. Because of the extensive wear, all that remains of the complicated zapodid tooth pattern are a few enamel islands on the tooth present, an M^1. Therefore, it is impossible to assign this form to a species. The tooth is similar in size to one of *Zapus princeps* J. A. Allen but is slightly narrower. The anteroposterior length is 1.4 mm, and the transverse width is 0.9 mm. The posterior border of the alveolus for the P^4 is present (Fig. 7E) and definitely relates this specimen to the genus *Zapus*.

Order CARNIVORA

Material.—Doby Springs local fauna, no. 38787, a medium-sized canine and no. 38788, a small deciduous carnassial

Remarks.—These were the only carnivore teeth found. The small deciduous carnassial belonged to an animal the size of a skunk. The canine compares most closely in size and shape with the upper canine of an otter.

Order PROBOSCIDEA

Family ELEPHANTIDAE

Mammuthus sp.

Mammuthus BURNETT, 1830, Quart. Jour. Sci., London, v. 28, p. 352

Material.—Doby Springs local fauna (?), no. 38798, tooth plates

Remarks.—Several plates of a mammoth tooth were recovered half a mile west in sec. 9, T. 27 N., R. 24 W., in deposits approximately contemporaneous with the fossil beds. These beds are in one of the red zones. Part of a proboscidian rib was found a quarter of a mile east of the above-mentioned location in the NE¼ of that section.

Order LAGOMORPHA

Family LEPORIDAE

Lepus sp.

Material.—Doby Springs local fauna, no. 38792, upper cheek tooth

Remarks.—A single tooth of a hare belonging to the genus *Lepus* was recovered. It is the size of a tooth of the whitetail jackrabbit.

Order ARTIODACTYLA

Family CAMELIDAE

Camelops sp.

Camelops LEIDY, 1854, Acad. Nat. Sci. Proc., Philadelphia, v. 7, p. 172

Geologic distribution.—Nebraskan to Wisconsin

Habitat.—The Pleistocene camels probably lived in the upland prairie regions. They probably could tolerate cold weather.

Material.—Doby Springs local fauna, no. 38797, a left astragulus

Remarks.—The astragulus compares favorably in size and configuration with the astragulus of *Camelops*.

Family ANTILOCAPRIDAE

Material.—Doby Springs local fauna, no. 38789, a left P_4

Remarks.—The premolar is that of an advanced antilocaprid and probably belongs to the genus *Antilocapra*. The roots have begun to form. The height of the crown is 22.0 mm, the transverse width measured at the occlusal surface is 5.0 mm, and the anteroposterior length is 9.8 mm. The tooth has three well-developed enamel pits. The middle pit runs transverse to the anteroposterior length of the tooth, and the other two parallel it. The posterior pit is 3.5 mm deep, and the other two are 7.0 mm deep.

Family BOVIDAE

Bison cf. *B. latifrons* (Harlan)

(Fig. 7J–L)

Bison latifrons (Harlan). SKINNER and KAISEN, 1947, Am. Mus. Nat. Hist. Bull., v. 89, p. 131–256

Geologic distribution.—Illinoian to Sangamon

Habitat.—The giant bison probably lived either along permanent streams, which supported shrubs and some trees, or in well-wooded stream valleys, in which there were intermittent areas of tall-grass land (meadow) (Hibbard, 1955a, p. 199).

Material.—Doby Springs local fauna, no. 38796, parts of jaws, teeth, and skeleton

Remarks.—Fragments of a large bison skull and skeleton were found 100 feet west of locality 1 in the summer of 1956. Because no horn cores were recovered, positive identification is impossible. Three left lower teeth (Fig. 7J–L) measure as follows (all measurements taken at the occlusal surface): P_2—anteroposterior length 14.6 mm, transverse width, 10.3 mm; P_3—anteroposterior length, 22.7 mm, transverse width, 14.0 mm; M_3—anteroposterior length, 48.5 mm, transverse width across first loop, 21.6 mm, transverse width across middle loop, 20.3 mm, and transverse width across posterior loop, 9.9 mm. Marie Hopkins of Idaho State College measured jaws associated with known skulls of *Bison latifrons*. These measurements compare favorably with those of the lower jaw of the Doby Springs specimen. The measurements taken were the width, 35 mm below the alveolus for the M_3 under the anterior border of the second loop, and the width, 26 mm below the alveolus in the vicinity of M_1 and P_4. The measurements on the Doby Springs material were 38.0 mm and 26.0 mm respectively, and the Idaho State College specimens measure from 34.2 to 44.4 mm and 31.6 to 39.3 mm, respectively.

Order PERISSODACTYLA

Family EQUIDAE

Equus cf. *E. niobrarensis* Hay

(Fig. 7H,I)

Equus niobrarensis HAY, 1913, Proc. U. S. Nat. Mus., v. 44, p. 576

Geologic distribution.—Illinoian to Sangamon

Habitat.—This was one of the large horses that probably spent most of its life moving along the lowlands and lake shores. It probably fed on the meadow grasses and prairie grasses close to the lowlands.

Material.—Doby Springs local fauna, no. 38790, right M^3; no. 38791, left P_3; no. 38793, partial metapodial; no. 38795, astragulus and calcaneum

Remarks.—The right M^2 and left P_3 are of a size comparable to that of the type of *Equus niobrarensis* Hay. The anteroposterior length of M^2 is 27.0 mm, the transverse enamel width is 25.0 mm, and the length of the protocone is 13.2 mm. The M^2 (Fig. 7I) has a wide postprotoconal valley;

the prefossette has a notched anterior border and is reduced anteriorly. A pli cabalin fold is present. The P_2 (Fig. 7H), from an old individual, has an anteroposterior length of 29.0 mm and a transverse width of 18.8 mm. It has a rather complicated pli hypoconid as is seen in *E. niobrarensis* (Hay, 1913, p. 579, Fig. 20). The metapodial, astragulus, and calcaneum belong to a large horse the size of *E. niobrarensis*.

Equus sp.

Material.—Doby Springs local fauna, no. 33334, front and hind phalange

Remarks.—The left M_3 belonged to a horse in the size range of *Equus conversidens* Owen (following Hibbard, 1955b, p. 61). The anteroposterior length is 26.2 mm, and the greatest transverse width is 12.3 mm. The tooth is smaller than the same tooth in a horse the size of *E. niobrarensis* Hay and belongs to a smaller group of horses.

The front and hind phalanges do not fit into the size range of either *E. niobrarensis* or the small horse. They belong to a horse of intermediate size. The front phalange is 79.5 mm long, 40.7 mm across the proximal end, and 37.1 mm across the distal end. The hind phalange is 68.4 mm long, approximately 42.0 mm across the proximal end, and 36.0 mm across the distal end.

PALEOECOLOGY

Fossil Zoogeography

The deposit from which this local fauna was collected represents a small part of the Illinoian glacial age. The climatic conditions during any glacial age certainly varied with the advance and retreat of the glaciers. The ecological conditions at the time the fauna lived are inferred from the fossils recovered and from the sediments in which they occurred. The writer assumes that the fossil forms had the habitat preferences of living members of their species group and that the smaller vertebrates had rather restricted habitats and are thus the best indices of climatic conditions.

Most of the fauna is extant, and therefore the writer could plot the present-day ranges of these forms and derive an area of common range overlap (Fig. 8). The distribution of the mammals is taken from Burt and Grossenheider (1952). Only the mammals were considered in preparation of this map. Smith (1954, p. 288) plotted the distribution of the fishes from the Berends local fauna, Beaver County, Oklahoma, and from his work on the Doby Springs fish material (1958) concludes that the area of common distribution of the species represented is the same for both faunas (Smith, oral communication). The author is indebted to C. L. Smith for permission to reproduce his map here (Fig. 9). The two maps show that the area of common distribution for the mammals, in the southeast corner of North Dakota and the northeast corner of South Dakota, lies within the western edge of the fish distribution.

On the basis of the forms represented four communities that contributed forms to the fauna are recognizable: the lake and marsh border community, the rich lowland meadow community, the shrub and tree community, and the upland prairie community. In addition to the forms belonging to these communities, larger animals having a wide range would, at one time or another, occur in each community.

LAKE AND MARSH BORDER COMMUNITY: Those animals living strictly in the marsh and lake environment are *Probythinella*, all the fishes, the salamanders, possibly the frogs and turtle, the muskrat, the northern water shrew, and the giant beaver. All these forms or their relatives live today in or around shallow to fairly deep small lakes in the northern part of the United States. *Ondatra* and *Sorex palustris* (muskrat and northern water shrew) spend most of their life in the water and are excellent swimmers. The fishes identified by Smith indicate an upstream pond environment (Smith, 1958, p. 180).

LOWLAND MEADOW COMMUNITY: The animals living in the rich lowland meadow are the arctic shrew (*Sorex arcticus*), the meadow vole (*Microtus*), the plains pocket gopher (*Geomys*), and probably the large horse (*Equus niobrarensis*) and the bison. The high abundance of the arctic shrew and the meadow vole indicates that a cool, moist, low meadow with good grass cover surrounded a large part of the lake. The plains pocket gopher prefers areas of extensive soil development but normally not the damp lowland area. It probably lived more on the valley slope than in the meadow. The large horse and bison probably spent most of their life near areas of water but also would have been found on the uplands.

SHRUB AND TREE COMMUNITY: Both the jumping mouse (*Zapus*) and the shorttail shrew (*Blarina*) indicate a presence of some trees in the area. *Blarina* today favors areas having some leaf mold. Their presence in the

fauna does not require forests, but a few stands of trees seem to be indicated. The jumping mouse also lives in meadows, and it probably was a form that lived in both communities. community. Both ground squirrels prefer the upland sandy soil areas to the low meadows. The western pocket gopher lives today on the dry prairies, whereas the plains pocket gopher

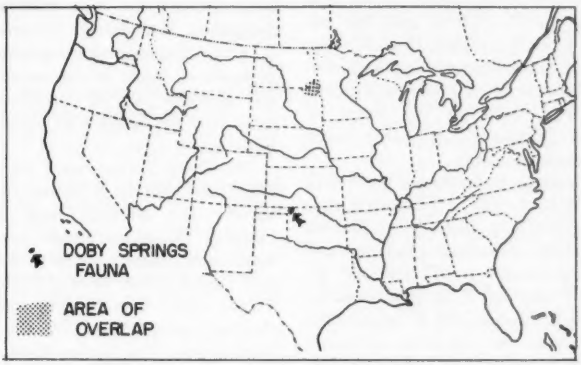


FIGURE 8.—DISTRIBUTION OVERLAP OF EXTANT MAMMALS OF THE DOBY SPRINGS LOCAL FAUNA

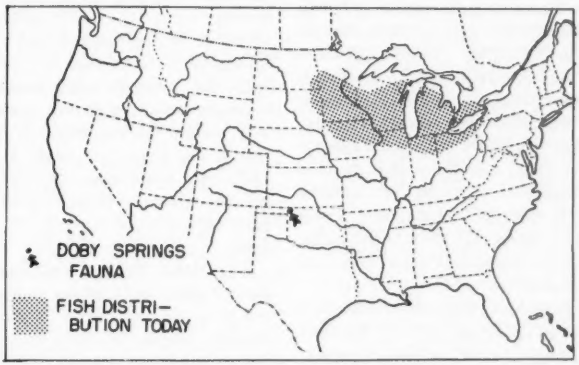


FIGURE 9.—PRESENT DISTRIBUTION OVERLAP OF ALL SPECIES OF FISHES OF THE DOBY SPRINGS LOCAL FAUNA
(After Smith, 1954)

The recovery of only one specimen of *Zapus* and one of *Blarina* indicates that they probably were not abundant forms in the immediate vicinity of the lake.

UPLAND PRAIRIE COMMUNITY: Included in the upland prairie community are the masked shrew (*Sorex cinereus*), the ground squirrels (*Citellus*), the western pocket gopher (*Thomomys*), the northern grasshopper mouse (*Onychomys*), and the hare (*Lepus*). The masked shrew lives in spots of rich vegetation in the prairies. It is probably transitional between this community and the meadow (*Geomys*) lives in areas of better soil development. *Thomomys* burrows in dry sandy soils. The presence of the grasshopper mouse indicates that there were probably some spots of open ground with sparse vegetative cover.

COMPOSITE COMMUNITY FORMS: The grazing animals were probably wide-ranging forms and were not confined to any one community. The grazing animals (mammoth, camel, pronghorn, and horses) lived in the areas of grass, coming to the lake for water. The smaller horses probably spent more time on the upland than did the larger ones.

Climate

The determination of the climate at the time the Doby Springs local fauna lived is based on the climate of the area in which the extant forms are all living today, in the area of southeastern North Dakota and the adjacent corner of South Dakota (Fig. 8). The limits of the area are determined by the southwestern border of the range of *Sorex palustris* and the eastern border of the range of *Onychomys* and *Thomomys*. All other forms occur in and around this area.

The mean annual temperature in North Dakota is 42° F., whereas in Oklahoma it is 57° F. (Visher, 1954, map 3). In North Dakota the average difference between January and July mean temperatures is 60° and 65° F., whereas it is 40° and 50° F. in Oklahoma for the same period (Visher, 1954, map 328). The precipitation in the two areas as shown on map 492 (Visher, 1954) is about equal at about 25 inches per year, of which 70 per cent occurs in the months of April through September. Because of the lower temperature in North Dakota, less evaporation occurs there than in Oklahoma. Thornthwaite (1948) classified climates on the interrelationships between temperature, precipitation, and evaporation. Following his classification, Harper County, Oklahoma, today has a dry subhumid climate belonging to the third mesothermal with little or no water surplus ($C_1B_2'db_2'$). The area of range overlap in North Dakota and South Dakota has a dry subhumid climate belonging to the fourth mesothermal with little or no water surplus ($C_1B_1'db_1'$).

The Doby Springs local fauna lived in Harper County, Oklahoma, during part of Illinoian time with a climate similar to that of southeastern North Dakota and northeastern South Dakota today. On the basis of the fauna and the alternation of silts and clays the writer assumes that a small lake existed around which there were marshes, rich meadows, and some trees and shrubs. Farther away from the lake were some well-drained upland prairies. The average annual temperature was probably about 42° F. with about 25 inches of precipitation per year, of which 75 per cent fell during the summer months.

Age and Correlation

The Doby Springs local fauna is considered Illinoian in age as determined by stratigraphic position, faunal evidence, and correlation with other faunas. The lake deposits in which the fauna was recovered overlie the Kansan-Yarmouth Crooked Creek formation and thus must be younger than Yarmouth. The caliche developed in the top of the lake beds is considered to have formed during the Sangamon interglacial. Therefore, the fauna is post-Yarmouth and pre-Sangamon. The presence of forms in the fauna now living in the northern part of the United States indicates that the fauna is of glacial age and thus Illinoian.

The Doby Springs local fauna is correlated with the other local faunas of Illinoian age, namely the Berends local fauna, Beaver County, Oklahoma, and the Butler Spring local fauna, Meade County, Kansas (Hibbard and Taylor, 1960; Smith, 1958; Taylor, 1943; Tihen, 1955). The Berends local fauna offers the best comparison with the Doby Springs local fauna. The faunas have five species of mammals in common. Smith (1958) has compared the fishes from the Berends local fauna, the Butler Spring local fauna, and the Doby Springs local fauna. He considered them as Illinoian in age and found that the Doby Springs local fauna had five species in common with the Berends local fauna and three in common with the Butler Spring local fauna. The mammals common to all three local faunas are abundant forms in glacial faunas. Sufficient vertebrate material was not recovered from the Butler Spring local fauna to allow comparison with the exception of the fishes.

The Doby Springs local fauna is more advanced than the Cudahy fauna (Kansan) and the Borchers local fauna (Yarmouth) of Meade County, Kansas, in that a greater percentage of extant species is present. Only 2 of 16 species in the Cudahy fauna are extant, and only 2 of 23 in the Borchers local fauna (Hibbard, 1949a). The Doby Springs local fauna contains 14 extant species of 24. Comparison with the Sangamon Jinglebob and Cragin Quarry local faunas (Hibbard, 1949a; 1955a; Hibbard and Taylor, 1960) is difficult because of the presence of southern forms in these faunas that are absent in the Doby Springs local fauna. The Wisconsin Jones local fauna, Meade County, Kansas (Hibbard, 1949a), is made up almost entirely of extant species (11 of 13). The earlier age of the Doby Springs local fauna is further substantiated by the presence of two extinct *Peromyscus* species, whereas in the Jones local fauna a *Peromyscus* belonging to the *P. maniculatus* group is present.

latus or *leucopus* group was recovered (Hibbard, 1940). Extinct species of *Peromyscus* are also recovered from Sangamon faunas (Hibbard, 1958), and therefore the Recent species of *Peromyscus* apparently were not differentiated until Wisconsin time.

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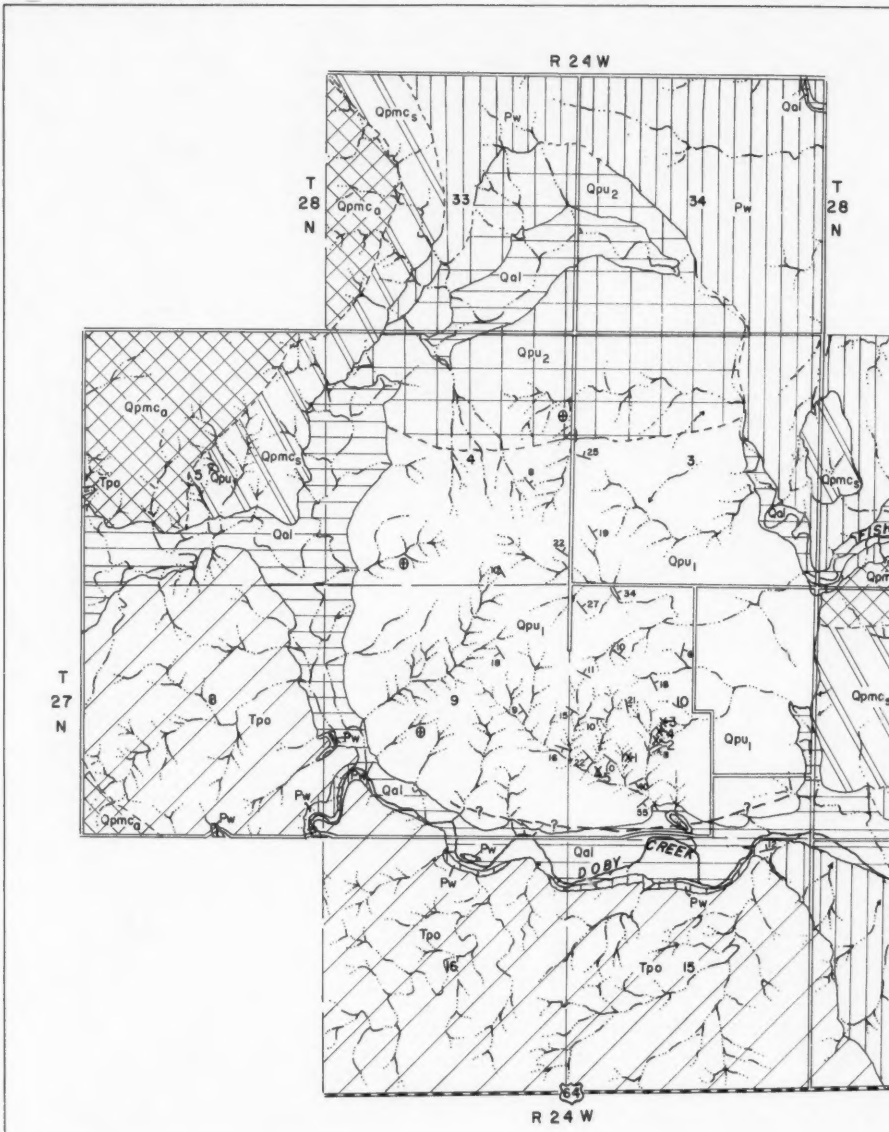
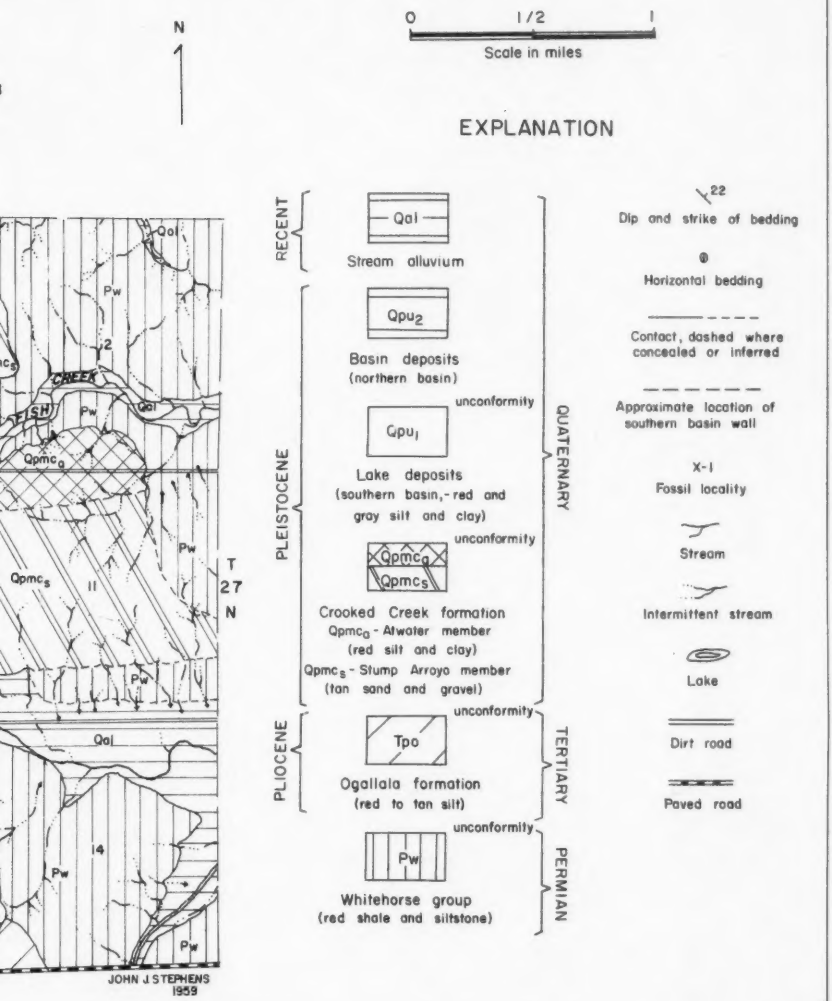


GEOLOGIC MAP OF DOBY SPRINGS ARE



AREA, HARPER COUNTY, OKLAHOMA

Short Notes

BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
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SALINE LAKES AND DRILL-HOLE BRINES, McMURDO SOUND, ANTARCTICA

BY DONALD G. BALL AND ROBERT L. NICHOLS

ABSTRACT

A 50-foot hole was drilled in bed rock at Marble Point, McMurdo Sound, Antarctica. Although the bottom of the hole was −3°F. several inches of brine were found there. It had moved down joints and into the drill-hole from an intermittent saline lake.

Saline Lakes

A shallow lake a few hundred yards long, which dries up during the summer months, is located about 2 miles northwest of Marble Point, McMurdo Sound, Antarctica. It is fed by meltwater from the snow which falls into and drifts into its drainage basin. It was also fed, at an earlier time, by meltwater from the nearby Wilson Piedmont Glacier. When the lake dries up a flat bottom is exposed. Test pits indicate that in places there is as much as 3 feet of fine-grained lake sediments which rest on till and perhaps on bedrock.

A trench was cut in these sediments by a bulldozer during the early part of February 1958. Ground water flowed into the trench, and a small pool about 1 foot deep was formed, the surface of which was nearly 2 feet below the top of the lake sediments. The water in this trench had a distinct taste. An analysis by the Water Analysis Laboratory of Metcalf and Eddy, Engineers, Boston, Mass., is as follows:

pH	7.5
Alkalinity as $CaCO_3$ (ppm)	230
Hardness as $CaCO_3$ (ppm)	3400
Magnesium as Mg (ppm)	390
Calcium as Ca (ppm)	710
Chlorides as Cl (ppm)	3300

The analysis shows that the concentration of salts is very much less than that of sea water. The principal dissolved salts are probably NaCl, $MgCl_2$, and $CaCO_3$.

There are at least three possible sources for the salts. The very high concentration of chlorides suggests that the dissolved material, in part at least, was blown in fine spray from the ocean about half a mile away. In sea water there is approximately 3 times as much magnesium as calcium (Clarke, 1920, p. 123). The water in the trench, however, contains nearly twice as much dissolved calcium as magnesium. The high concentration of calcium is easily accounted for, however, as the country rock of the area is a calcite marble. An efflorescence is common in the Marble Point area, on till and on outwash, beach, and lake deposits. It usually forms a very thin veneer but in some places it is a crust a few millimeters thick. It is friable, soft, soluble, and has a strong taste. Prof. Cornelius S. Hurlbut, Jr., of Harvard University made an X-ray analysis of it and found it to be approximately 50 per cent halite (NaCl) and 50 per cent thenardite (Na_2SO_4). The Water Analysis Laboratory of Metcalf and Eddy analyzed this material also. The efflorescence was washed off several pieces of soil with a stream of distilled water, and an analysis of the solution is as follows:

	mg per 100 ml
Dissolved solids	192
Calcium as Ca	1.76
Magnesium as Mg	0.86
Sulphates as SO_4	59.0
Chlorides as Cl	50.0

The halite (NaCl) was undoubtedly derived from the ocean. The thenardite (Na_2SO_4) was probably formed by the interaction of H_2SO_4 resulting from the oxidation of pyrite, on minerals containing sodium. It seems likely

that the efflorescence is at least a minor source for the dissolved salts.

During the full-bodied stage of the lake the concentration of the dissolved salts is not great. After all the snow in the lake basin is melted the lake water slowly evaporates, and the concentration of dissolved salts slowly increases. With the advance of the cold season the ground water in the lake deposits freezes from the top downward, and a more and more concentrated brine is formed at depth. The eutectic temperature of pure NaCl dissolved in water is approximately −6.0°F., and the eutectic temperature of pure $MgCl_2$ dissolved in water is about −31.0°F. The presence of the $MgCl_2$ will, however, depress the eutectic point of the NaCl, and the presence of the NaCl will depress the eutectic point of the $MgCl_2$. Sea water is said to be a complete solid at about −65°F. As the mean monthly temperatures for this area during the winter months are probaby between −10°F. and −15°F. (Simpson, 1919, p. 81) some of the brine will probably remain unfrozen for most and probably for all of the year.

No study was made of the hydrologic regime of the lake. It is believed, however, that water flows out of the lake across a low threshold during the late spring and early summer. At this time, however, the depth of thaw in the deposits beneath the lake is presumably not great, and the brine will not be removed because it is probably trapped in frozen ground. If, on the other hand, the sediments containing the brine are completely thawed while the lake is still overflowing it seems unlikely that much of the brine will be removed from the lake basin because it is in the sediments beneath the lake, it has a high specific gravity, and diffusion of the brine into the lake waters would be slow. In either case the dissolved salts are not easily removed, and they may have been accumulating for many years.

Because of the proximity of the ocean, the possibility that the dissolved salts were in part derived from relic sea water suggests itself. Raised beaches as much as 61 feet above sea level are found about half a mile from the lake. The lake, however, is 100–200 feet above sea level. If any of the dissolved material were derived from relic sea water it would have had to come from an earlier more extensive submergence before the last glaciation.

The precipitation in the Marble Point area is mainly or entirely in the solid state, and it is probably about 2 inches (water equivalent) a year (Personal communication, from Commander Henry C. Stephens, U.S.N.).

No measurements were made to determine how much of the snow cover is dissipated by evaporation and how much by melting. The presence of sizable nieves penitentes, plowshare snow surfaces, and sun cups indicates that a considerable percentage of the snow cover is dissipated by evaporation. Where the snow cover is thin it may be entirely dissipated by evaporation before the free-air temperatures get high enough to melt the snow. Where the snow cover is thick it lingers long enough into the spring and summer months to be in part melted. Perhaps not more than 1 inch of water is available for runoff. This extreme aridity is of course an important factor in the accumulation of the salts.

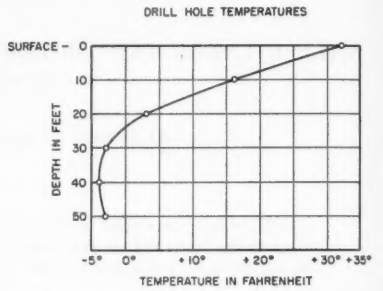

FIGURE 1.—CURVE SHOWING THE DISTRIBUTION OF TEMPERATURE IN THE DRILL HOLE

Drill-Hole Brines

In the early part of February 1958, a hole was drilled in marble to a depth of 50 feet with a pneumatic drill. The temperatures at various depths in the hole were as follows (Fig. 1):

Depth (feet)	*Temperature* (°F)
at surface	+32
10	+16
20	+3
30	−3
40	−4
50	−3

The mean annual temperature at Hut Point, Cape Royds, and Cape Evans, McMurdo Sound, for the 5-year interval 1902, 1903, 1908–1909, 1911, and 1912, was +0.7°F. (Simpson, 1919, p. 81). These localities are approximately 50 miles from the Marble Point

area, and the mean annual temperature in the Marble Point area is, therefore, not far from 0°F. An analysis of the temperature curve for the drill hole (Fig. 1) and a consideration of the mean annual temperature of the area suggests

Metcalf and Eddy, Boston, Mass., who supervised the drilling of the hole, reports that at a depth of approximately 12 feet an earth-filled joint was encountered. Apparently a saline solution which had its origin in the lake sedi-

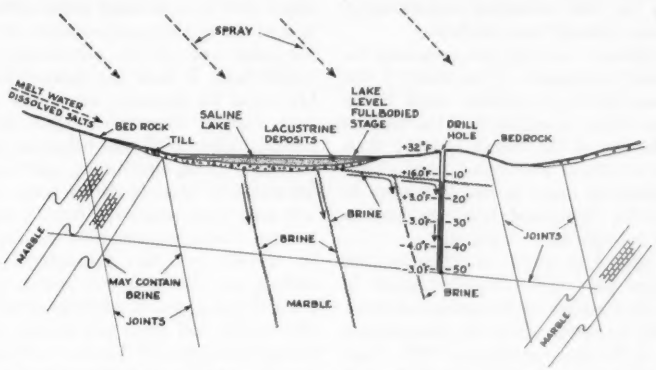

FIGURE 2.—DIAGRAMMATIC SKETCH SHOWING THE SALINE LAKE AND THE BRINE IN THE JOINTS AND DRILL HOLE

The brine entered the drill hole from the slightly inclined joint which intersects it.

that the −4°F. encountered at a depth of 40 feet in the drill hole is very close to the temperature below the zone of seasonal temperature changes in this area.

Several inches of a very concentrated brine were found at the bottom of the drill hole where the temperature was −3.0°F. The Water Analysis Laboratory of Metcalf and Eddy analyzed the brine as follows:

pH	6.4
Alkalinity as $CaCO_3$ (ppm)	5,500
Hardness as $CaCO_3$ (ppm)	128,000
Calcium as Ca (ppm)	27,200
Magnesium as Mg (ppm)	14,400
Chlorides as Cl (ppm)	137,000

The analysis shows that the concentration of the brine is very much greater than that of sea water. The collar of the hole is approximately 120 feet from the pool in the bulldozed trench from which the water sample was collected, and it is much closer to the edge of the lake deposits. It is approximately 4 feet above the top of the lake sediments and about 5½ feet above the surface of the pool. The proximity of the lake deposits to the collar of the drill hole and the chemical similarity of the brine and the lake waters indicate that the brine in the drill hole was derived from the saline lake waters. Mr. J. J. Scheuren, Jr., ments entered the drill hole at this point, ran down into the hole, and reached the bottom (Fig. 2). The saline solution probably reached the bottom of the drill hole shortly after the hole was completed. As it moved down the hole its temperature was lowered, ice crystallized out, the brine became more salty but remained liquid even at the bottom as its eutectic temperature was below −3°F. If the temperature at the bottom of the hole was 1°F. warmer than it was 10 feet higher at a depth of 40 feet, and the temperature measurements indicate that it was, the downward movement of the brine which was probably above 10°F. when it entered the drill hole was probably responsible.

The geologic, climatologic, and physical conditions and processes which enabled the brine to move downward to a depth of at least 12 feet in marble are of interest. The brine undoubtedly moved down the joint which cuts the marble and which intersects the drill hole at about 12 feet. If this joint had sufficient width and if it were open and continuous from the bottom of the mantle rock down to the drill hole the brine could very easily have reached the drill hole. The geologic history of the area suggests, however, that the joint was ice-filled rather than air-filled.

As is well known, more or less vertical cracks

sometimes form in frozen ground when the ground contracts because of a significant lowering of its temperatures. The ice in the joint, however, was probably too narrow to have contracted sufficiently to have formed a crack. It appears, therefore, that a crack or passageway for the downward movement of the brine was probably not available.

No passageways, however, are necessary for the downward movement of the brine; it can melt its way down an ice-filled crack if the saline thread is long enough so that the temperature of the top of the thread is lower than that at the bottom of the thread. Such a temperature difference could be due to a wave of cold advancing downward from the surface because of the approach of winter.

Imagine a pool or thread of saline solution in the upper part of an ice-filled crack in marble which is buried by lacustrine sediments and till. Let us assume that the temperature at the top of the saline solution is 30°F., that the marble at this depth is also at 30°F., that the temperature of the free air is 30°F., that the temperature both in the ice-filled crack and in the marble decreases downward until a continuous uniform temperature of −5°F. is reached, and let us also assume that at the existing temperatures and salinities the solution and the ice have the same vapor pressure and are therefore in equilibrium.

If the temperature of the free air drops from 30°F. to 15°F. then the temperature at the top of the saline thread will drop, and the upper part of the saline thread will freeze. This will destroy the equilibrium between the saline solution and the ice below it, and some of the ice will melt until equilibrium for the new conditions is established. The new solution will be more concentrated than the old because not so much ice will melt immediately below the thread as was formed in the upper part of the thread, it will be smaller in volume, and in the process it will have melted its way a certain distance downward in the ice-filled crack.

With further lowering of the free-air temperature the saline thread will continue to move downward and to increase in concentration until the brine solidifies because its eutectic temperature is reached or until the brine reaches an environment of uniform temperature or until the crack pinches out. A repetition of the process the following season would drive another thread downward. The process which enables the brine to melt its way downward in the ice-filled crack would also allow it to move initially from the mantle rock down into the upper part of the ice-filled crack.

Still another mechanism will also enable brine to eat its way downward in an ice-filled crack. Imagine a small bubble of brine in the upper part of an ice-filled crack. Let us assume that at the existing temperatures and salinities the brine and the ice surrounding it are in equilibrium. If now the temperature of the brine and ice increases, because of a wave of heat moving downward from above, but remains below 32°F., the brine and the ice will no longer be in equilibrium, and some of the ice will melt. More of the ice below the bubble will melt than above the bubble, because the volume of water formed from the melting of the ice will be less than the volume of the ice melted, and the brine will tend to draw away from the ice above it. Melting of the ice above the bubble will stop, but melting of the ice below the bubble will continue until equilibrium is again established. If now the temperature of the ice and brine decreases because of a wave of cold advancing rapidly downward, ice will crystallize out of the brine until equilibrium for the new temperature is reached. If as much ice should crystallize out at the top of the brine as at the bottom the brine bubble in the ice-filled crack would move downward because of this alternation of temperature. If on the other hand more ice crystallized out at the top of the brine than at the bottom the brine bubble in the ice-filled crack would move downward. More water will not crystallize at the bottom of the brine than at the top because the wave of cold advances downward. A continuation of temperature changes will result in a continuous downward movement of the brine in the ice-filled crack. When the brine reaches its eutectic temperature or the zone of uniform temperature no more downward movement will take place.

Acknowledgments

The observations on which this note is based were made in February 1958. The data were provided by the Commanding Officer, U. S. Naval Construction Battalion Reconnaissance Unit.

The writers are happy to acknowledge the help they received from Commander Henry C. Stephens, U. S. N., Mr. J. J. Scheuren, Jr., and Mr. A. F. Ferullo of Metcalf and Eddy, Engineers, Boston, Mass., and Professors

Robert D. Eddy and Charles E. Messer of Tufts University.

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BRECCIA BLOCKS (MISSISSIPPIAN) OF THE WELCOME SPRING AREA, SOUTHWEST UTAH

BY E. F. COOK

Abstract

Breccia blocks of Mississippian limestone, resting on gravel and sand of probable late Pliocene or Pleistocene age, previously interpreted as remnants of a Tertiary thrust sheet, probably were not pushed but slipped from the adjacent mountain slopes, where they had been parts of a thrust sheet.

In a recess of the western front of the Beaver Dam Mountains in the southwestern portion of Washington County, Utah (Figs. 1, 2) about 25 masses of brecciated Mississippian limestone rest on poorly sorted and unconsolidated gravel and sand. The breccia blocks range in area from about 1 acre to 160 acres and in thickness from 50 to about 200 feet. These breccia blocks have been regarded as faulted and eroded remnants of a Medial or Late Tertiary thrust sheet and represent the only known evidence for such recent thrusting in southwest Utah.

The breccia block area is just west of Welcome Spring, 2½ miles northwest of Castle Cliff on U. S. Highway 91. All the blocks are found within a fairly narrow vertical range, 3400–4400 feet above sea level. Although some of the blocks, especially those nearer the mountain front, are inclined toward the mountains, the altitude of individual blocks diminishes in irregular step fashion southwesterly away from the mountain front. None of the blocks stands more than 300 feet above the general level of the adjacent slopes.

East and north of the breccia block area the Beaver Dam Mountains rise abruptly to crestal elevations of 5800 feet (east) and 6800 feet (north).

North of Castle Cliff the Beaver Dam Mountains consist of a resistant mass of Precambrian schist and gneiss overlain on the northeast and east by several thousand feet of Paleozoic sedimentary rocks that dip away from the Precambrian core.

Castle Cliff itself is part of a low-lying sheet of Pennsylvanian Callville limestone thrust onto Precambrian and lower Paleozoic rocks. The Castle Cliff thrust, first recognized and named by Dobbin (1939, p. 129–131), can be traced along a narrow outcrop, except for a short covered interval, to a place about 4 miles northwest of Castle Cliff where it ends against a high-angle fault.

On the hill east of the breccia-block area is a west-dipping klippe of Mississippian Redwall limestone (it may have some older strata in its basal portion) resting on Precambrian rocks. The Mississippian klippe, the higher part of which is known as Sheep Horn Knoll, is distinguished from the Mississippian breccia blocks by (1) its larger size (twice as large as the biggest breccia block); (2) its relative lack of brecciation; (3) the fact that it lies on Precambrian rocks instead of Cenozoic gravel; and (4) its greater elevation.

Available exposures indicate that poorly sorted gravel of diverse rock types underlies the breccia blocks. In one good exposure near the mountain front the gravel is composed largely or entirely of fragments of upper Paleozoic rocks; it contains no recognizable Precambrian material despite the nearness of a great mass of Precambrian rock. Farther from the mountains the gravel contains clasts of sedimentary rocks of Paleozoic to Eocene(?) age and volcanic rocks of Oligocene and Miocene age; it appears to overlie, unconformably, well-bedded sand and silt of probable Pliocene age. In the best exposure of the base of a block the bedding in the gravel is parallel to the base of the block, and the gravel shows no deformation or disturbance that might be attributed to the emplacement of the overlying breccia.

Dobbin called the gravel under the blocks a conglomerate of possibly Miocene age and thought that the blocks might represent a post-Miocene(?) thrust. He wrote (1939, p. 131),

> "The large size of the limestone blocks and their structural relations to the underlying conglomerate . . . suggest that they are remnants of a once continuous sheet of Mississippian strata that was thrust over Miocene (?) alluvial and playa deposits. . . . Many of the cobbles in the limestone conglomerate were derived from the Callville limestone, probably from the overriding block of the Castle Cliff thrust."

Reber (1952, p. 107) also regarded the brecciated Mississippian blocks as remnants of a once-continuous sheet that had moved eastward over Miocene(?) sediments.

FIGURE 1.—LOCATION MAP

Five facts suggest that the breccia blocks may be slide blocks rather than klippes: (1) the geometric relations in the breccia-block area, which provide a hill from which the blocks could have slid; (2) the thinness and brecciated nature of the blocks which would probably render them incapable of transmitting any considerable stress; (3) the presence of flowage structures and the lack of flat movement planes within the blocks; (4) grooves on the underside of a block that are parallel to the inferred direction of sliding; and (5) the presence, on the hill above the breccia blocks, of an unbrecciated klippe of the same rock, the basal shear zone of which could have facilitated sliding.

The breccia blocks lie below some of the steepest slopes of the west front of the Beaver Dam Mountains. The structure of the Sheep Horn Knoll klippe (Fig. 2) suggests that it may have been bent, after its emplacement, by uplift of the block north of it (although the bending could represent molding of an erosion thrust to the shape of the overridden surface); such uplift may have been accompanied by slope steepening sufficient to initiate sliding of other portions of the thrust sheet of which the Sheep Horn Knoll klippe is a remnant.

Unless they have been greatly—and rather uniformly—reduced in thickness by erosion after emplacement, the breccia blocks, according to the thrust hypothesis, represent a thin sheet of limestone that would seem, especially in view of its brecciated condition, incapable of transmitting the stress necessary to push it into position. On the other hand, mechanical weakness is no handicap to sliding.

The relatively flat shear planes caused by differential translation, which one expects to find in the basal portion of a thrust sheet, are not found in the breccia blocks. Swirling masses of brecciated chert within the blocks suggest that they were emplaced in part by flowage.

Shallow parallel grooves in the base of the best-exposed breccia block, presumably formed at the time of emplacement, are oriented in the direction of inferred sliding. In other words the direction of the grooves is northeast, approximately parallel to the bisectrix of the angle or embayment of the mountain front in which the blocks are concentrated.

The Sheep Horn Knoll klippe provides an argument against thrusting as well as a suggestion of the source of the breccia blocks. If the klippe and the breccia blocks—in their present positions—were once part of the same thrust sheet, their internal structural features should be similar. But the klippe is not broken like the breccia blocks, and from sparse exposures the basal portion appears to be a shear zone with planar slip surfaces, similar to the zone at the base of the Castle Cliff thrust sheet where exposed west of Welcome Spring. The Sheep Horn Knoll klippe may remain in its thrust position today only because normal faulting has dropped it against Precambrian rock on its downhill side, into a position whence it was impossible to slide (Fig. 2). The breccia blocks may represent other portions of the same thrust sheet that did slide.

Parenthetically, the Sheep Horn Knoll klippe may itself be a slide block, moved by gravity from the topographically higher Mississippian outcrops to the northeast. The origin of this klippe, however, is not materially involved in the question of whether the breccia blocks slid or were pushed into their present positions.

Gravel beneath a breccia block in the SW¼ sec. 35, T. 42 S., R. 19 W., contains cobbles and boulders of volcanic rocks of Miocene age. Identified volcanic boulders and cobbles are from formations that crop out only to the north, none within less than 15 miles of the breccia-block area. The youngest volcanic formation represented in identified clasts is the Cove Mountain of H. R. Blank (unpublished Ph.D. thesis, Univ. Wash.), which is probably equivalent in part to the Page Ranch formation, the upper unit of which has a zircon age of 19 m.y. (Mackin, 1960, p. 98).

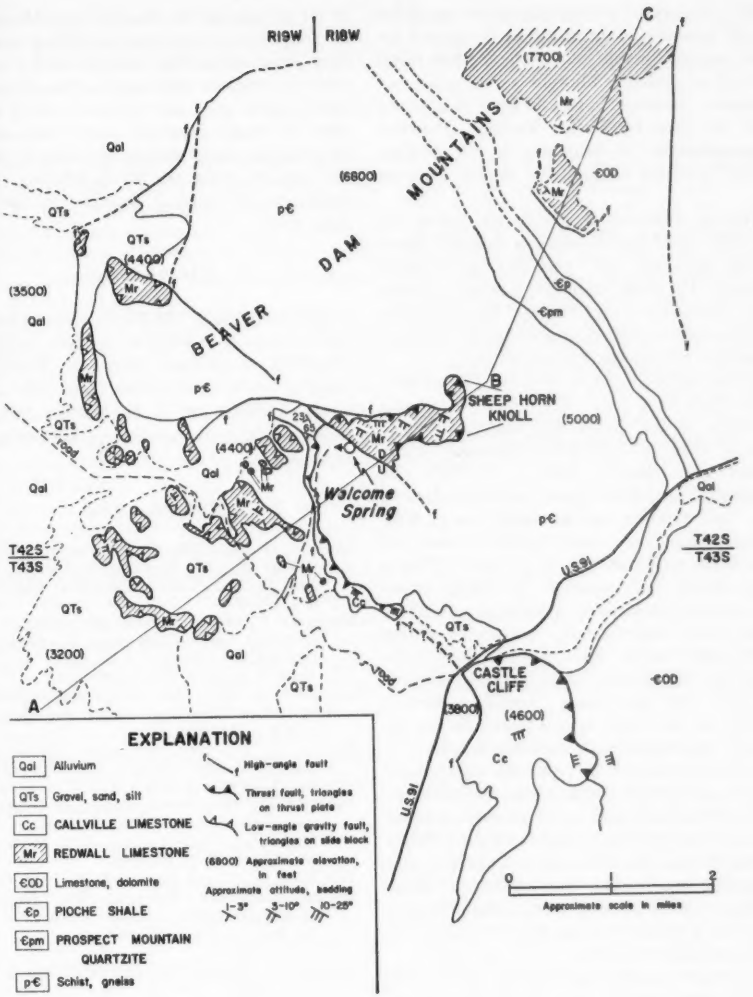

FIGURE 2.—GEOLOGY OF THE WELCOME SPRING AREA, WASHINGTON COUNTY, UTAH

Geology mapped on air photos of Soil Conservation Service, Virgin River area (1938); base from Utah State Highway Department map

More important in determining the age of the gravel, however, is the fact that the gravel appears unconformably to overlie pinkish sand and silt of probable Pliocene age. Similar fine sediments elsewhere in Beaver Dam Wash have been identified by C. M. Tschanz (Personal communication) as belonging to the Muddy Creek formation of probable medial Pliocene age.

The age relations thus date the gravel beneath the cited breccia block as definitely post-medial Miocene, and probably post-medial Pliocene. The evidence of fast-flowing and relatively abundant water recorded by the thickness, coarseness, and poor sorting of the gravel suggests a Pleistocene age assignment.

Coarse monolithologic breccias of probable slide origin have been described from several localities in the Basin and Range province (Grose, 1959, p. 1935–1937; Longwell, 1951; Woodford and Harriss, 1928, p. 279–283). The interfingering of these breccia masses with ordinary fan deposits points clearly to a sedimentary-landslide origin, with intermittent downslope movement of large blocks during fanglomerate accumulation. Longwell called such slow or intermittent movement a "chronic slide." Most authors have interpreted sliding of this nature as the result of uplift, slope steepening, and subsequent erosion along steep faults.

Unlike the previously described breccia masses, the Welcome Spring blocks do not appear to interfinger with ordinary fan deposits or to grade laterally or vertically into bedded breccia of the same composition. The inference drawn is that the blocks were all emplaced during a relatively brief episode of sliding. If the sliding did take place more or less abruptly, the triggering mechanism was probably either sudden slope steepening of considerable magnitude or a climatic change that rather quickly decreased slope stability.

In light of the Pleistocene(?) age of the sliding, it seems probable that Pleistocene rainfall was at least a contributory factor in the slipping of the klippes off the Beaver Dam Mountains. Although the primary cause of sliding was probably slope steepening brought about by tectonic or intrusive deformation, the advent of a pluvial cycle may have induced sliding sooner than it would otherwise have occurred, by creating an unprecedented hydrostatic head in the pendant slabs and by facilitating the development of internal mobility in the basal shear zone.

Acknowledgments

This brief study was made in the summer of 1959 as part of a mapping project incident to compiling a geologic map of Washington County, Utah, for the Utah State Land Board. C. M. Tschanz and J. H. Mackin aided materially in the formulation of the ideas here expressed.

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GEOLOGY OF THE DUNCHURCH AREA, ONTARIO, CANADA

BY W. C. LACY

Geologic mapping in the vicinity of Dunchurch, Ontario, during the summers of 1940 and 1941 was interrupted by the war, but recent papers and accompanying maps (Thomson, 1956; Friedman, 1957) showing the Dunchurch area as unmapped have stimulated compilation of the field and laboratory data.

The Dunchurch area, located about 20 miles northeast of Parry Sound, Ontario, constitutes the northeast portion of the Sundridge map sheet bounded by 45° 30′ to 45° 45′ N. latitude and 79° 45′ to 80° 00′ W. longitude (Fig. 1). This includes portions of McKellar, Spence, Ferguson, Hagerman, Croft, McKenzie, and Ferrie townships.

The area lies in the westernmost portion of the Grenville subprovince and contains the typical marble, metasedimentary rocks, and intrusive rocks that characterize the Grenville series farther east. Rock units strike generally north to north-northeast and dip steeply eastward. The structure is characterized by broad domal features, generally elongated northnortheast, separated by tight isoclinally folded synclinal troughs of metasedimentary rocks and marble. The domal areas are occupied by feldspathized banded gneisses, coarsely layered hypersthene granite and hypersthene diorite, gabbro, or anorthosite. Faulting is minor.

Grenville Series

Two distinct variations are represented in the Grenville series in the Dunchurch area: (1) an upper (?) unit composed of metasedimentary rocks and marble, and (2) a lower (?) unit, the Ahmic gneiss, consisting of well-banded microcline-biotite and garnet amphibolite gneisses and quartzite. Metamorphism reached the intensity of the amphibolite and granulite facies, then underwent retrogression.

The Whitestone Lake metasedimentary rocks form tightly folded belts about half a mile wide composed of rusty-weathering gneiss with marble. The marble forms low persistent ridges in soil-covered valleys and is best exposed along lake shores where the marble appears to have influenced shape and distribution of the lakes. Thickness of the marble beds ranges from only a few inches enclosed in rusty-weathering gneisses to local thicknesses in excess of 1000 feet, but averaging 50 to 75 feet. There may be as many as five separate marble beds within the marble–metasedimentary unit. The marble beds are generally somewhat magnesian, and metamorphism resulted in formation of minerals characteristic of the amphibolite facies which have suffered later retrograde metamorphism evidenced by pseudomorphic forms. The metamorphic minerals occur as disseminated grains and as folded, concentric, or parallel aggregates within the marble. They include:

Metamorphic minerals	*Retrograde equivalents*
Hornblende-parasite (β= 1.660–1.708)	
Tremolite	
Garnet (almandite, pyrope)	Chlorite
Diopside-augite (β = 1.702–1.715)	Serpentine, scapolite
	Brucite, chondrodite
Periclase	Serpentine, scapolite
Olivine (forsterite)	Scapolite (mizzonite)
Anorthite-andesine (An_{30-65})	Phlogopite

The rusty-weathering gneisses associated with the marble are generally strongly deformed. They consist of thinly bedded quartzites, amphibolites, and feldspar-hornblende paragneisses. The amphibolites all have gneissic structure but vary considerably in width and composition. The composition of amphibolite from four localities is given in Table 1.

Quartzites occur as discontinuous lenses in the rusty-weathering gneiss. They consist of relatively pure aggregates of well-sorted and rounded quartz grains with minor orthoclase and microcline pods up to a few inches in length developed along bedding planes.

The Ahmic gneiss appears to lie just below the Whitestone Lake metasedimentary rocks and forms a thick sequence of light-colored well-bedded gneisses believed to have been of sedimentary and/or volcanic origin. Their dip is generally concordant with that of the marble-metasedimentary unit. The gneiss ranges from mica schist to microcline feldspar gneiss. Simple

pegmatites are abundant in this unit as dikes following fractures or as lenticular pods or streaks paralleling the foliation. Fluorite is a common accessory mineral. A measured section

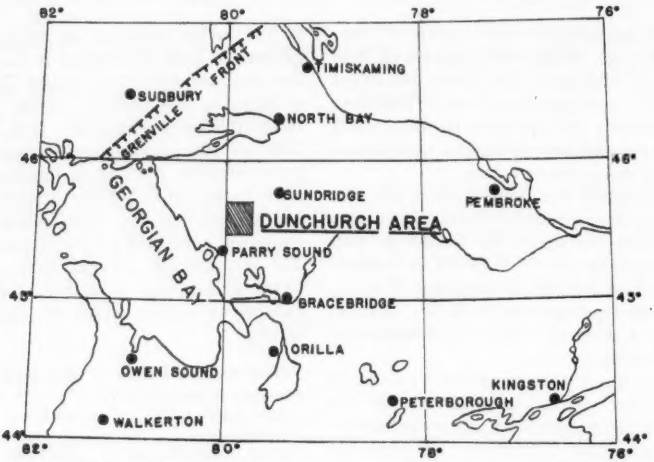

FIGURE 1.—INDEX MAP OF CENTRAL ONTARIO SHOWING LOCATION OF THE DUNCHURCH AREA

across this unit northeast from Ahmic Harbour illustrates the variation within the unit:

Feet	Fault
600	Gray garnet-biotite-microcline augen gneiss
70	Gray biotite-microcline augen gneiss with pegmatite segregations
630	Well-bedded pink and gray hornblende-biotite-garnet gneiss
210	Well-bedded pink hornblende-feldspar gneiss with microcline augen
50	Massive gray feldspar-biotite gneiss
400	Well-bedded pink and gray feldspar-amphibole gneiss with texture ranging from saccharoidal to coarsely crystalline
150	Massive gray feldspar gneiss
150	Massive pink microcline gneiss
140	Well-bedded pink and gray biotite-amphibole-feldspar gneiss
360	Massive saccharoidal microcline gneiss with red hematite-bearing spots and amphibole and pegmatite segregations
60	Well-bedded pink and gray gneiss
350	Massive gray amphibole-biotite gneiss, highly contorted with streaks of microcline and pegmatite segregations
580	Massive fine-grained pink to gray biotite-hornblende gneiss and many concordant pegmatite segregations
3750	

Dome-forming Rock Series

The central portions of the domes, which dominate the structural picture of the Dunchurch area, range in composition from granite to anorthosite.

The granite areas are gradational into the Ahmic gneiss and appear to have been derived from the gneiss by a process of feldspathization. The granites occur in irregularly shaped masses characterized by an abundance of pegmatites. Microtextures show evidence of granulation followed by regrowth of feldspars producing the granitic appearance. Typical material collected half a mile southwest of Ahmic Harbour gave the following modal analysis (per cent):

Quartz	33.5
Microcline	41.5
Oligoclase (An_{15})	13.4
Biotite	4.0
Hornblende	7.5

McKellar gneiss consists of massive coarsely banded gray-green to pink gneissic rock containing persistent dark bands that range from a few inches to several hundred feet in thickness. It occurs in definite elongated domal structures and in many places is separated from the Grenville metasedimentary rocks by the marble zone. Pegmatitic phases are small and sparsely distributed. The McKellar gneiss ranges in composition from granite to hypersthene diorite as shown in the modal analyses of Table 2.

The gabbro and anorthosite are closely related in appearance and in composition. They form elongate, domal structures concordant

TABLE 1.—COMPOSITION OF AMPHIBOLITE

	1	2	3	4
	%	%	%	%
Garnet (almandite)	20	20	—	5
Augite	20	—	—	—
Hypersthene	2	—	—	—
Hornblende	20	5	48	33
Biotite-chlorite	—	—	8	—
Andesine	35	1	41	50
Microcline	—	—	—	5
Magnetite	2	73	—	—
Apatite	1	—	1	—
Corundum	1	1	—	—
Quartz	—	—	3	5
Pyrite	—	—	—	2

Chemical Analysis of Sample No. 1 (per cent)
(Analysis by F. A. Gonyer)

SiO_2	46.96
TiO_2	2.72
Al_2O_3	14.16
Fe_2O_3	3.00
FeO	12.37
MnO	0.15
MgO	6.02
CaO	10.60
Na_2O	2.94
K_2O	0.48
H_2O+	0.57
P_2O_5	0.14
V_2O_5	0.13
	100.24

1. 1 mile north of North Whitestone Lake
2. 4½ miles southeast of Dunchurch
3. 2 miles west of Ahmic Harbour
4. 3 miles west of Ahmic Harbour

TABLE 2.—MODAL ANALYSES OF MCKELLAR GNEISS

	1	2	3	4	5	6	7
Quartz	24.6	24.5	8.8	—	—	—	—
Orthoclase	59.6	39.0	—	—	—	—	—
Plagioclase	12.9	29.0	72.6	50.7	48.8	71.1	44.3
Biotite	1.5	—	—	—	—	1.8	—
Hornblende	—	—	13.6	28.1	2.9	25.0	15.3
Augite	—	—	—	13.9	27.6	—	8.4
Hypersthene	—	7.2	4.7	4.7	15.6	—	25.2
Magnetite	1.4	—	—	2.9	4.9	1.2	6.5
Apatite	—	—	.1	—	—	.9	—

Chemical Analysis of Sample No. 7
(Analysis by F. A. Gonyer)

SiO_2	48.90
TiO_2	2.26
Al_2O_3	13.90
Fe_2O_3	4.42
FeO	8.68
MnO	.09
MgO	6.47
CaO	10.56
Na_2O	3.28
K_2O	.58
H_2O+	.46
P_2O_5	.11
V_2O_5	.09
SO_3	.02
	99.82

1. NW shore of Minerva Lake
2. McKellar
3. Owl Lake
4. Ryan's Lake
5. Middle River
6. NW shore of Minerva Lake
7. McKellar

with the adjacent metasedimentary rocks and frequently lie immediately adjacent to the marble zone. The gabbro domes are aligned along a north trend passing to the east of North Whitestone Lake. Three domes separated by the Whitestone Lake metasedimentary rocks are recognized. The rock is greenish-gray and shows rough banding. It becomes gneissic close to the contact with the metasedimentary rocks. Composition ranges between the following limits:

	Per cent
Quartz	0–15
Orthoclase	0–5
Plagioclase (An_{20-55})	70–90
Augite	0–15
Hornblende	2–20
Biotite	0–5
Garnet	0–2

Mortar structure is characteristic, and alteration of the feldspars to scapolite is common.

The anorthosite forms a large elongate body 4 miles wide and 10 miles long extending north-northeast from Shawanaga Lake. It forms generally massive, white outcrops. Near its contact with the metasedimentary rocks it becomes roughly banded with narrow bands and streaks richer in ferromagnesian minerals. North and northeast of Shawanaga Lake it contains knots of ferromagnesian mineral concentrations, giving it a spotted appearance. This spotted anorthosite is richer in scapolite, pyrite, and ferromagnesian minerals than is the bulk of the anorthosite. Scapolite is abundant throughout the contact zone of the anorthosite; it replaces the plagioclase, particularly where it was granulated and formed mortar structures. Modal analyses of two specimens of the anorthosite from the eastern contact area are given in Table 3.

TABLE 3.—MODAL ANALYSES OF ANORTHOSITE

	1	2
Quartz	—	0.5
Plagioclase	83.2 (An_{58})	69.4 (An_{54})
Scapolite	13.4 (Me_{75})	14.5 (Me_{70})
Hornblende	3.1	9.7
Garnet (almandite)	0.3	3.8
Sphene	—	1.9

1. North shore Shawanaga Lake
2. Northwest shore Lorimer Lake

Structure

The structure in the Dunchurch area is similar to that mapped by Adams and Barlow (1910) in the Haliburton and Bancroft area. It is characterized by broad elongate domal structures trending north to northeast with a steep eastward dip and occupied by massive to gneissic rocks ranging from granite to anorthosite. The interdomal areas are occupied by tight isoclinally folded synclinorium belts of metasedimentary rocks that are not appreciably metamorphosed by the adjacent dome-forming rocks. Buddington (1959) proposes that these domical plutons of the Grenville subprovince represent a major form of granite emplacement in the catazone.

In the metasedimentary rocks foliation is well developed and follows the bedding. Lineation is pronounced and plunges generally away from the domal centers.

Faulting is of minor importance in the area. An east-northeast to east-trending fault cuts off the Ahmic gneiss just south of Ahmic Harbour; elsewhere, minor breaks appear to have influenced drainage along east-northeast and north-northeast directions, but there has been no appreciable displacement.

Metamorphism of Metasediments

Intensity of metamorphism of the metasedimentary rocks in the Dunchurch area is relatively uniform in developing the granulite-amphibolite facies. There is no apparent change in intensity of metamorphism adjacent to the dome-forming rocks, but there is a greater concentration of pegmatites and feldspathization of the metasedimentary rocks along the projected strike of the domal structures where they terminate against the metasedimentary rocks or where the metasedimentary rocks are involved in domal structures. This might be interpreted as indicating upward migration of hyperfusible components from the dome-forming rocks into the metasedimentary rocks that were subjected to tension forces over the crests of the domal structures, but with little lateral migration of these components into the compressed interdomal areas.

Interpretations

Microtextures showing crushing and regrowth of the minerals in the dome-forming rocks suggest that these rocks assumed their present position by flow in a nearly solid state. Lack of reaction with or metamorphism of the marble–metasedimentary unit immediately adjacent to the dome-forming rocks is indicative of either equilibrium between the rock types or other conditions not favoring reaction. In the crest portions of the dome structures, where tension conditions existed, there resulted a concentration of hyperfusible components that effected granitization of the metasedimentary rocks.

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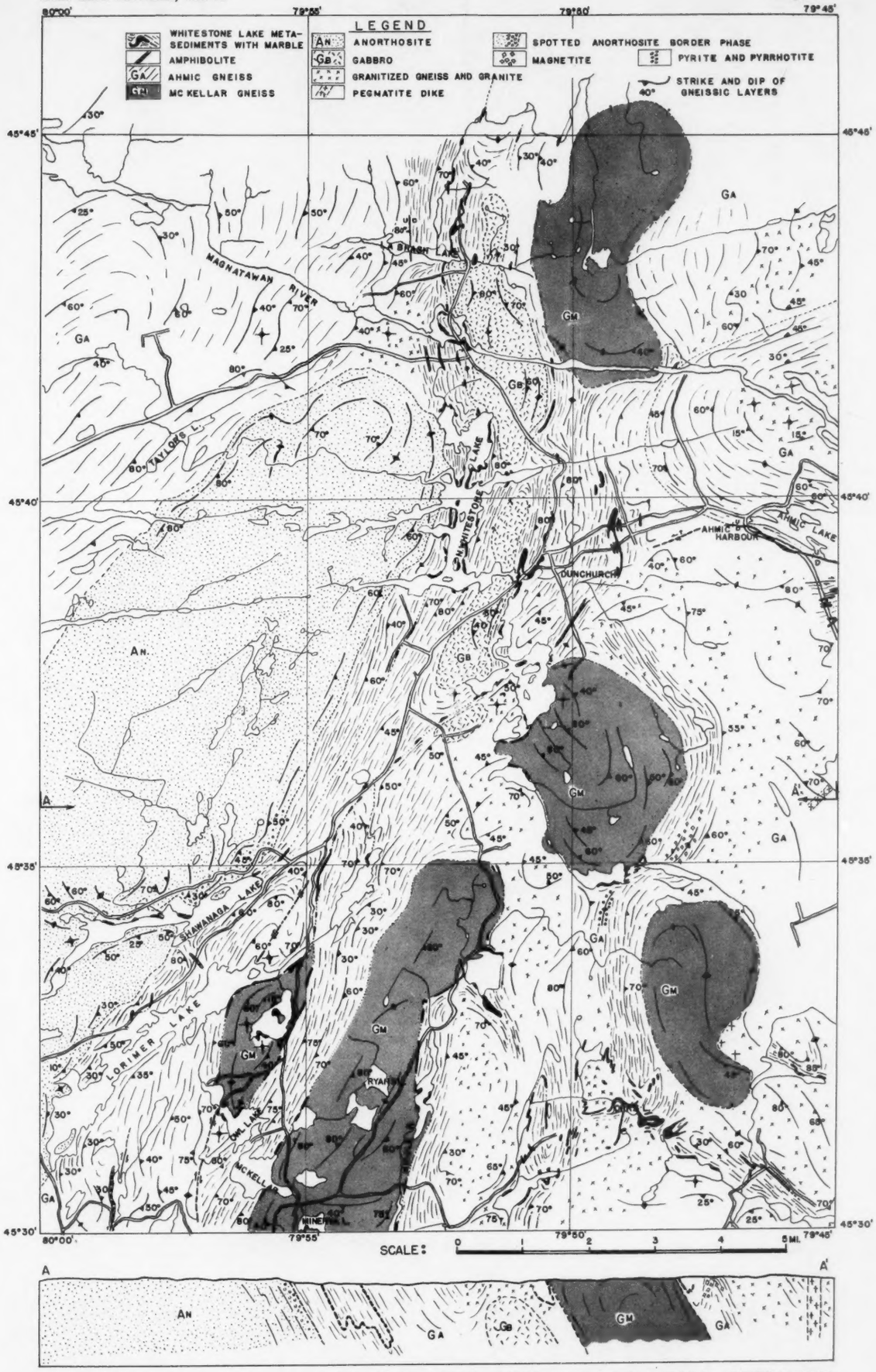

GEOLOGIC MAP AND SECTION OF THE DUNCHURCH AREA, ONTARIO

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GRAPHICAL STATISTICAL ANALYSIS OF FRACTURE PATTERNS IN ROCK ENCOUNTERED IN ENGINEERING PROJECTS

BY AARO E. AHO

Abstract

The directional competence of badly fractured rock in engineering foundations can be analyzed by plotting the various types of fractures on a Schmidt equal-area projection and combining the data suitably. Interpretation of the results may increase safety factors and efficiency in grouting, strengthening devices, design, and tunnelling.

In some engineering projects bedrock is very badly fractured and causes concern especially where nonvertical loads or seepage and its complications are involved, as in dam sites and in tunnels. In many such cases the problem can be obviously solved by further excavation, by dowels, and by similar devices to take advantage of greater bearing surface or stronger rock, or by grouting to strengthen or leak-proof the rock. These procedures are usually guided by good common principles, and adequate treatment provides a large enough safety factor. However, more precise knowledge of the nature of such a fractured condition could increase the efficiency of these procedures, thereby tending to increase the safety factor and to decrease the costs. This paper calls attention to an application of a graphical statistical geologic method (Bucher, 1944; Fairbairn, 1949; Aho, 1957) by which the dominant orientations and nature of fractures can be determined for such purposes regardless of their apparent complexity and abundance.

In this well-known method the strike and dip of numerous fractures are measured. The types of fractures recorded can be separated into useful categories such as:

(1) Faults—large prominent fractures usually filled with gouge. The width of gouge should be indicated.
(2) Slickensided fractures—smaller, more discontinuous fractures with small gouge seams or slick walls
(3) Dikes or veins—(often with faults or slickensided fractures)
(4) Joints—minor fractures with clean, rough, or irregular walls.

For adequate statistical representation 100 readings or more are usually necessary, although in simple cases a smaller number may be representative.

In addition to fractures, any detailed geology should be plotted on a plan of the area, and the prominent fault or fracture zones should be shown on this in their actual locations. However, the main mass of strikes and dips can be recorded without reference to their exact location except in a foundation site where the degree of fracturing, rock type, or apparent orientation of fractures varies in different areas. The readings of fracture orientations should be grouped for separate analysis of different localities, if necessary, to show this variation in pattern. Unless the fracture surfaces are highly sinuous, an average reading on representatively scattered fractures is often sufficient. Highly sinuous fractures may require averaging of readings. The observer should neither consciously avoid favoring any one trend or type of fracture, nor should he over-compensate, but he should try to distribute his readings on the basis of prominence and numerical abundance unless every fracture is measured. However, even the best observer may show minor unconscious bias as a result of various factors beyond his control.

In the above procedure it is assumed that the rock is essentially homogeneous, as would be the case for most granitic rocks. Other structural features such as platiness or foliation (cleavage), stratification, banding, folds, etc., may complicate the picture and often predominate over the fracture pattern. These features would be mapped conventionally, but if they are very complex a similar statistical approach can be applied to their analysis.

The poles of the 100 or more readings from apparently randomly oriented fractures are now plotted on the lower hemisphere of a Schmidt equal-area projection, resulting in a scatter diagram.

The different types of planes—faults, slickensided fractures, dikes, veins, or joints—can be plotted on separate sheets with different colors

or symbols of dots, crosses, circles, etc., for the poles. The overall pattern can be seen by superimposing the different sheets of tracing paper or by combining them, according to type or locality.

If desired, the scatter diagram can be contoured to aid in visualization of density of poles (Fairbairn, 1949, Ch. 21). The contouring gives an approximate numerical value to the relative abundance of fractures of any particular range of orientation and produces a pattern that sometimes can be visualized and compared more easily than a scatter diagram.

Interpretation will vary depending on specific data and related geology. For this reason conventional geologic mapping should precede or accompany any such fracture analysis. Limited exposures of bedrock seldom give complete data, so a true three-dimensional picture cannot be obtained from exposures in one plane only, especially if prominent fractures lie parallel to that plane.

A paper by the writer (Aho, 1957) shows an application of this type of approach in mineral exploration.

It is interesting to note that, if the approximate coefficient of friction on the fracture surfaces were reliably known, perhaps some quantitative measure of the competence or strength of the rock might be calculated for comparative purposes by integrating the coefficients and orientations in relation to stress. However, indeterminate factors such as irregularities and discontinuities in the fractures would probably invalidate such a mathematical procedure in most cases.

In spite of its limitations, this statistical method of analyzing fractures has the following obvious engineering applications:

(1) Grouting can be done most effectively by drilling holes oriented at steep angles to the planes of dominant fracturing.

(2) Design (*e.g.*, in dams) can be modified to avoid placing maximum stress on the planes of dominant weakness.

(3) Dowels, anchors, and other strengthening devices can be placed or oriented for maximum strength.

(4) Excavation or tunnelling can be planned to take advantage of planes of weakness, to avoid disturbing the rock, or to expose the rock in the best manner.

(5) Interpretation along with other geologic work may lead to further useful conclusions; other applications of this type of approach may appear in specific cases.

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ON THE ANTIQUITY OF "SEDIMENTATION" AND HYDROLOGY

(WITH SOME MORAL CONCLUSIONS)

BY PAUL D. KRYNINE

Abstract

The ancient Greek philosophers held very modern views on sedimentation. Democritus discovered shape sorting of pebbles on beaches; Aristotle worked out the mechanical aspects of pebble abrasion; Plato produced truly post-Huttonian descriptions of erosion and deposition around Athens, and he also brought forth, 1900 years before Palissy, the correct, pluvial, theory for the origin of springs and rivers.

Since some leading modern physicists have rediscovered the fruitfulness of certain basic principles of early Greek scientific thought, perhaps geologists should also look into it.

The Facts

The systematic study of sediments as such (*i.e.*, sedimentary geology, sedimentary petrography, or "sedimentation"[1] for short) is a recent scientific enterprise, dating back to 1920 or perhaps even to 1935 or so. No mention whatsoever is made of this subject in the various books on the History of Geology, at least in the English language (Adams, Geikie, Lyell, Zittel).

Thus, it is interesting to find that among the ancient Greeks "sedimentation" is one of the few aspects of geology that was treated in an entirely modern way.

In support of this, I am offering three quotations, from Democritus, Aristotle, and Plato, respectively. To the best of my knowledge these excerpts, although naturally known to classicists, have never been published in any book on the history of geology, or on the history of science in general (Sarton, Dampier, Clagett, Duhem, Farrington, and quite a few others).

Democritus (Demokritos), fl. 420 B.C., was the associate of Leucippus (Leukippos), the "shadowy" but real inventor of atomism. Democritus is considered to be the organizer of the atomistic theory of matter, and his intellectual powers were perhaps only slightly inferior to those of Plato and Aristotle. Although the full text of the enormous Democritic corpus (between 61 and 70 titles, Freeman, 1953, p. 297–299) has been lost, a sufficient number of fragments and later commentaries, from Aristotle on, have survived to inspire a gigantic literature on atomism, ranging from Epicurus to Schrödinger. Diels, in his famous collection of the *Fragments of pre-Socratic philosophers*, gives numerous quotations from Democritus and also doxographic comments on the great atomist.

An excerpt from Diels' Fragment 68 (55)B 164 is translated by Freeman (1948, p. 107) as follows:

"On beaches . . . with the motion of the waves, oval pebbles are driven to the same place as oval, and round to round. . . ."[2]

The Nahm translation (Nahm, 1947, p. 189, item 128, and p. 218, item 164, actually translated by Gordon H. Clark) is a trifle different:

"For even on the seashore the similar pebbles are seen in the same places, here the spherical pebbles, there the long ones. In sifting also the same shapes collect in the same places so that the beans and the pulse fall apart from each other."

This is not a bit bad; as a matter of fact it is good enough to be given verbatim in a lecture on sedimentary shape sorting and hydraulic equivalence in an advanced course on Sedimentary Petrology taught in the year 1959 A.D.

But, in addition to effective shape sorting, the form of pebbles in a sedimentary deposit may be

[1] The term "sedimentation" is a popular catch all. Properly used, it should be restricted to processes dealing with the *immediate* origin of sediments. The term is mostly used in this sense in the present note.

[2] The more extended text of this fragment in the Freeman translation reads: "So it is with inanimate things, as one can see with sieving of seeds and with the pebbles on beaches. In the former, through the circulation of the sieve, beans are separated and ranged with beans, barley grains with barley and wheat with wheat; in the latter with the motion of the wave, oval pebbles are driven to the same place as oval, and round to round as if the similarity in these things had a sort of power over them which had brought them together."

modified by wear during transport. An ingenious description of the possible mechanism of pebble abrasion, based upon the principle of the lever, is given by Aristotle in *Mechanica* (Ch. 15, 852b 29–853a 4). The authorship of *Mechanica* is somewhat in doubt, as it is not certain whether this minor work was written by the Master himself or by one of his pupils. The passage reads as follows:

"Why is it that the so-called pebbles found on beaches are round, though they are originally formed from stones and shells which are elongated in shape? Is it because objects whose outer surfaces are far removed from their middle point are borne along more quickly by the movements to which they are subjected? The middle of such objects acts as the centre and the distance thence to the exterior becomes the radius, and a longer radius always describes a greater circle than a shorter radius when the force which moves them is equal. An object which traverses a greater space in the same time travels more quickly, and objects which travel more quickly from an equal distance strike harder against other objects, and the more they strike the more they are themselves struck. It follows, therefore, that objects in which the distance from the middle to the exterior is greater always become broken, and in this process they must necessarily become round. So in the case of pebbles, because the sea moves and they move with it, the result is that they are always in motion, and as they roll about, they come into collision with other objects; and it is their extremities which are necessarily most affected."

I must admit that this dynamic approach to abrasion took me somewhat by surprise: perhaps through mental inertia, I used to think in static terms, namely that the inherent fragility of an elongated particle was greater than that of an equant one.

The third quotation is from Plato (429–348 B.C.) who is freely acknowledged as a supreme philosopher, as one of the greatest masters of prose that ever lived, and as no mean mathematician. As a physical scientist, however, Plato has not fared so well at the hands of some nineteenth century-style "moderns": too much idealism, mysticism, and the like. This opinion may be due to a literal or uncritical interpretation of Plato's cosmology in the *Timaeus*. Obviously these critics of Plato have not studied geology, or have not read the *Critias*, or have done neither.

The unfinished *Critias*, probably the penultimate of Plato's dialogues, written when Plato was perhaps over 70, contains an extraordinarily modern geologic passage. This is a description (*Critias* 110E to 111D and a sentence from 112A) contrasting the geomorphology and sedimentation of "Pre-Historic" Athens (*ca.* 9000 years before Plato) and "modern" or Platonic Athens (*ca.* 350 B.C.).

In a somewhat abbreviated form, freed from purely geographic and historic elements, this passage reads as follows, in the Bury translation[3]:

"... all other lands were surpassed by ours in goodness of soil .... And of its goodness a strong proof is this: what is now left of our soil rivals any other in being all-productive and abundant; ... and at that period (*note: i.e.*, 9000 years before Plato) it produced these things in vast quantity. How, then, is this statement plausible and what residue of the land then existing serves to confirm its truth? The whole of the land lies like a promontory jutting out from the rest of the continent far into the sea; and all the cup of the sea round about it is, as it happens, of a great depth. Consequently, ... during the 9000 years ... the soil which has kept breaking away from the high lands during these ages ... forms no pile of sediment[4] worth mentioning, as in other regions, but keeps sliding away ceaselessly, and disappearing in the deep.[5] And, just as happens in small islands, what now remains compared with what then existed is like the skeleton of a sick man, all the fat and soft earth having wasted away, and only the bare framework of the land being left. But at that epoch the country was unimpaired, and for its mountains it had high and arable hills, and in place of the "moorlands" (*note: i.e.*, stony soils or "phelleus"—φελλεύς) as they are now called it contained plains full of rich soil; and it had much forestlands in its mountains, of which there are visible signs even to this day ....Moreover it was enriched by the yearly rains from Zeus, which were not lost to it, as now, by flowing from the bare land into the sea; but the soil it had was deep, and therein it received the water, storing it up in the retentive loamy soil; and by drawing off into the hollows from the heights the water that was there absorbed, it provided all the various districts with abundant supplies of spring waters and streams, whereof the shrines which still remain even now, at the spots where the fountains formerly existed, are signs which testify that our present description of the land is true."

Immediately following (*Critias* 111E and 112A–B), Plato describes the probable change

[3] The 1929 Bury translation (part of the so-called Loeb classical series) is preferred to the 1871 Jowett (so called Oxford) translation which makes good reading but, in general, is somewhat less accurate. This applies specifically to the *Critias* and, according to the classicists, to the rest of Plato as well. It should be noted that most of the popular editions of Plato are from the Jowett translation.

A good bit of the *Critias* passage referred to has also been well translated by Burnet (p. 91).

[4] Translated by Burnet as "silt" and by Jowett as "settlement of the earth flowing down from the mountains."

[5] Plato mercifully refrained from inventing the allegedly omnipresent and omnipotent turbidity currents, so beloved by certain modern geologists, to account for this phenomenon.

in climate, from an "evenly temperate" to one of strong (periodic?) rains (*i.e.*, present Mediterranean type) and comments again upon the catastrophic violence of some cloudbursts: "For as it is now, the action of a single night of extraordinary rain has crumbled it (*note*: *i.e.*, the site of the "ancient" Acropolis) away and made it bare of soil. . . ." Finally the latest dissection of the "ancient" Athenian flood plain ("... level on the top . . .") is described. These interesting portions, however, are so interlarded with history and contemporary geography that it would be burdensome to quote them in full, and they become unintelligible when fragmented out of context.

What have we here? First of all, an exceptionally accurate account (and the first one on record) of rainwash erosion and sedimentation under various climatic and physiographic conditions. It even includes a talk on the sad effects of deforestation!

The charming poetic style makes such an account not entirely suitable for inclusion in the frequently dry-as-dust exposition of many a modern geology textbook, but the facts are clear enough.

Second, an entirely correct origin is postulated for springs and rivers, complete with proofs acceptable to any uniformitarianist. This must be stressed, because all positivists, from 1600 on, have been taking Plato to pieces, not only for the allegedly "evil" influence of the *Timaeus* on science but also for his picturesque, although fanciful description of the Homeric Tartarus (*Phaedo*, 111–114). But this description is an ironical one! ("... a charming tale, well worth hearing . . . ", *Phaedo*, 110E). Tartarus is a parageological (and Dantesque) Hell of fire and water inside the earth from which volcanoes emanate, and into which rivers flow so as to come out again, roasting and half drowning the souls of the damned in the process.

Tartarus was firmly rejected by Aristotle (*Meteorologica*, Book 11.2, 355b 32–356a 33, with the blame put on Plato!). Nevertheless, Tartarus proved to have exceptional staying powers. Naturally it was used as a model for numerous medieval hells and purgatories.[6]

[6] One of the most remarkable of these is described by Charles Singer (1928). His figure 94 shows a curious picture of "The Sphere of the Earth", as conceived by Hildegard of Bingen, a medieval abbess, visionary, and lady scholar (ca. 1200). Hildegard's "Earth" contains an interesting mixed hydrothermal and fiery interior, connected with the surface, where (p. 222): "Souls suffer the torments of flame in one section, the torment of water in another . . . ."

Finally, as late as 1664 Athanasius Kircher in his *Mundus Subterraneus* (*See* Adams, Ch. XIII) refurbished Tartarus with a perfectly logical—but incorrect—scientific foundation by creating the theory of the Alembic to account for the origin of springs and rivers through underground distillation of ocean water. This direct linear descent of the Alembic from Tartarus has apparently escaped the attention of geologic historians.

It does not seem quite right to blame Plato—or Homer!—for the peculiar lucubrations produced by the medieval scholastics. In *Critias* Plato shows less of the famous irony which permeates his earlier works, and his geological description of Athens is perfectly straightforward. The origin of springs and rivers postulated in the *Critias* is based on acute observation and not on a tongue-in-check embellishment of Homeric myths as in the *Phaedo*.

This Platonic—and also modern—hydrologic pluvial explanation of the *Critias* was only partially adopted by Aristotle, who grafted on it the unnecessary condensation concept of the "mountain sponge"; it was restated by Vitruvius (*ca.* 25 B.C.), and then it was forgotten. Tartarus took over (with Plato as the author!).

The pluvial hypothesis was revived, somewhat hesitatingly, only 1850 years later, by Leonardo da Vinci (1500), it was clearly reformulated by Bernard Palissy in 1580, and definitely proven by Pierre Perrault in 1674, *i.e.*, almost simultaneously with Kircher's Alembic theory![7]

[7] *See*, for details, Aristotle's *Meteorologica* Book 1.13–14, 349b 2–351a 18; Mather and Mason, 1939, p. 20–23 for a verbatim translation of Perrault; Sarton, 1927, p. 223 for Vitruvius; the Palissy translation (p. 48, 53, 55, 56), and a general summary in Adams, 1938, Ch. XIII.

Leonardo da Vinci apparently held contradictory views on the subject. Adams (1938, p. 446), on the basis of a secondary Italian source (Lorenzo, 1920), credits Leonardo with the concept of circulation of rain and snow water from mountain tops downward, through underground "pervious" beds, either toward the surface or directly into the ocean. I was not able to consult the Lorenzo book, and had to content myself with the study of a somewhat selective primary source—*i.e.*, a translation of Leonardo's *Notebooks*, edited by MacCurdy (1938). A reading of the geology, physiography, and hydrology sections (185 pages) produced at least 17 passages endorsing underground circulation of oceanic water along subterranean "veins" toward the mountain tops. This is a modification of Tartarus (vol. 1, p. 329, 338, 345, 346, 359, 360, 365, 367; vol. 2, p. 21, 22, 23, 34, 65, 75, 103, 107, 121). The analogy is with a plant or with man: after "rupture" of these pressurized veins high in the mountains water "naturally" seeks its proper

The Conclusions

The "moral" of this geologic "rediscovery" of Democritus, Aristotle, and Plato is two-fold:

First, there is the supreme human irony of the situation itself. In a contradiction worthy of Socrates, Plato in two admirable dialogues proposes two conflicting hydrologic theories. One theory, the soundly scientific one, is not considered seriously by anybody excepting Aristotle and Vitruvius, and it has taken almost 2000 years for it to be re-established and accepted in modern times. The other theory, the fanciful one, is offered outright as a pleasant myth and is plainly labelled as such. It is immediately adopted by scholars, philosophers, and scientists, and Tartarus reigns unmolested for nearly 2 millenia. What is more, this is achieved over the opposition of Aristotle, no mean task indeed! The parallel with modern science—including geology—is obvious: the sensational, the mysterious, the infallible (and, we should add, the fashionable) always appeal more to human nature than do the sober, the austere, and the abstract. This first moral conclusion, although probably eternally valid, is nevertheless pessimistic or even cynical.

The second "moral" is more hopeful and constructive. Recently (1954) one of the most distinguished atomic physicists, Erwin Schrödinger,[8] found it necessary to go back to the pre-Socratic philosophers (including Democritus) in order to grasp the ultimate principles of basic scientific thought which had been lost by the wayside in today's ultra-specialization of physics and in the pseudo-sophisticated but trivial fragmentation of epistemology. As this is written (November 1959) another equally well-known physicist, Werner Heisenberg,[9] not only completely rehabilitates Plato but proclaims him, in fact, to be the patron saint of subnuclear physics.

Similarly, a modern mathematician and philosopher of the very first rank, the late A. N. Whitehead, had already rediscovered in 1920 the value of the *Timaeus* as a source of straight thinking in understanding the basic principles of spatio-temporal dynamism ("becoming"). I believe that in geology, where the problems are somewhat different from those of Schrödinger's and Heisenberg's quantum physics, but not too different from those of Whitehead's structures, an equally desirable removal of intellectual cobwebs can be obtained through a clear-headed restudy of Democritus, of Plato (including the *Timaeus*, of course!) and particularly of *all* the works of Aristotle (and not only of his *Meteorologica*, generally favored by historians of geology). Such a study would be worth undertaking.

"place" downward toward the ocean, just like a nosebleed (vol. 2, p. 107). An ingenious filtration mechanism (vol. 2, p. 62, 107) and an even more ingenious distillation process, by underground heat, fully anticipating the Alembic (vol. 2, p. 121), are suggested. In two other passages Leonardo denies the role of melting snow and rain water in the formation of rivers (vol. 1, p. 364, 368–369). In one spot only does he state that rivers are formed from "clouds" (vol. 2, p. 64), and in another, brilliant paragraph he proves, through the presence of chestnut leaves in a Sicilian spring, the derivation of this spring from the underground circulation of water coming from a lake (vol. 1, p. 364). There are four or five other, obscure, passages. The general effect is rather ambiguous, and, since these *Notebooks* cover 40 years, it is not certain what Leonardo's final opinion was.

This critique does not diminish by one iota the colossal and well-documented achievements of Leonardo in other geological fields: paleontology, physiography, sedimentation, etc.

[8] Nobel laureate in physics, 1933: Wave mechanics.

[9] Also Nobel laureate in physics, 1932: Quantum theory and the uncertainty principle.

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THICKNESS AND CONSOLIDATION OF DEEP-SEA SEDIMENTS: A DISCUSSION

BY G. H. DAVIS

Abstract

Hamilton (1959) concluded that in most sediments excess pore-water pressure is equal to zero—that is, the hydrostatic pressure is at atmospheric pressure. This note points out that in terrestrial environments the occurrence of artesian water (excess pore-water pressure) is commonplace and widespread and that such excess pressure is the source of energy for flowing water wells and for many flowing oil wells, "gushers," or "blowouts".

Hamilton (1959) has combined the disciplines of geology, geophysics, and soil mechanics most effectively in his stimulating paper on the thickness and consolidation of deep-sea sediments. Certain points, however, although not basic to his discussion, require further clarification.

In Appendix A (p. 1423) Hamilton states that "in normal situations the pressure induced at any level in a rock section does not include the weight of the overlying water as effective pressure on the mineral grains which would cause consolidation." Elsewhere he quotes Levorsen (1954) to the effect that in oil-field reservoirs pressures are usually hydrostatic and not the combined pressure induced by the weight of rock and water. Thus, Hamilton concludes that in most sediments excess pore-water pressure is equal to zero or, as he uses the term, "the hydrostatic case." There are several objections to these statements.

For one thing, Hamilton uses the term "hydrostatic" in a sense different from Levorsen. Whereas Hamilton uses it to mean a condition of zero excess pore-water pressure, as used by Levorsen (1954, p. 386) the term "hydrostatic" does not require that excess pore-water pressure be zero at the well or other point of measurement, but merely that the pressure be determined by the height of water above the confined surface at which the pressure is determined. Thus, the reference cited by Hamilton does not apply to the question under discussion. Furthermore, if normal (zero excess) pore-water pressure were the general case there would be no source of energy for flowing artesian water wells, and most flowing oil wells, "gushers," or "blowouts."

As the terms are used by hydrologists, "confined" or "artesian" water is ground water which is under sufficient pressure to rise above the level at which it is encountered in a well but which does not necessarily rise to or above the surface of the ground (American Geological Institute, 1957, p. 16). Where artesian pressure is sufficient to raise water in a well above the top of the zone of saturation or the water table, excess pore-water pressure exists. If the pressure is sufficient to raise the water above the land surface a flowing well results. In terrestrial environments the occurrence of artesian water (excess pore-water pressure) is not rare or unusual but rather is commonplace and widespread.

Meinzer (1928) reviewed the subject of compressibility and elasticity of artesian aquifers and concluded (p. 289) that artesian aquifers are all more or less compressible and elastic, and that they supply water not only by permanent reduction of porosity but also by temporary reduction that is restored when wells are shut down. The source of energy to raise artesian water and compress the aquifer materials is, of course, the weight of the overlying deposits and contained water, which is supported in part by the grains of the aquifer and in part by hydrostatic pressure. When the hydrostatic pressure in the aquifer is reduced, as by withdrawal of water from a well, a part of the support for the overlying load is removed, and the aquifer compacts in proportion to the increased stress.

More recently decline in artesian head or—to use soil-mechanics terminology—decline in pore-water pressure has been described as a cause of compaction of water-bearing deposits and resulting land subsidence in many areas (Tolman and Poland, 1940; Poland and Davis, 1956; Davis, 1958; Winslow and Doyel, 1954).

The writer does not mean to imply by the foregoing discussion that he disputes Hamilton's thesis regarding deep-sea sediments, which are

laid down in a vastly different environment from that of terrestrial water-bearing deposits. The intention is rather to clarify the question of the occurrence of excess pore-water pressure in terrestrial environments.

A further comment applies to Hamilton's admonition (1959, p. 1402) to geologists to follow soil-mechanics terminology in describing consolidation of sediments. As used in soil mechanics, *consolidation* is the process by which earth materials become more dense through application of load; *compaction* is reserved for artificial consolidation—for example, the use of sheepsfoot rollers in "compacting" road beds to reduce porosity.

Compaction and consolidation are defined in the Glossary of Geology and Related Sciences (American Geological Institute, 1957) as follows:

Compaction. Decrease in volume of sediments, as a result of compressive stress, usually resulting from continued deposition above them, but also from drying and other causes.

Consolidation. In geology any or all the processes whereby loose, soft, or liquid earth materials become firm and coherent.

This writer favors the retention of the geologic definitions of consolidation and compaction as given by the Glossary, not only on the grounds of prior use but also on the grounds of common usage, etymology, and convenience.

Both consolidation and compaction as defined above bear a closer resemblance to the general definitions as given by Webster's Dictionary (1959) than do the definitions cited by Hamilton. The Glossary definitions are preferable also on the basis of etymology. Consolidation is derived from the Latin *consolidare*, to make firm, from *solidus*, solid. Note that the Latin meaning applies to the character of the material, not the process by which it becomes firm. Compaction is derived from the Latin *compingere*, to join. In common English usage the word has come to imply closely or firmly united or packed particles, or having a small surface in proportion to contents or bulk (Webster).

On the grounds of convenience, also, the geological definitions are more useful to geologists than are those favored by workers in soil mechanics. For example, the compaction of sediment by drying could not be described in Hamilton's terminology either as consolidation or lithification. Thus, still more terms would be required to cover concepts easily disposed of in the geological definitions.

In short, established geological definitions for the terms consolidation and compaction should not be abandoned in favor of the arbitrary, specialized definitions employed in soils-engineering work unless the engineering terms can be shown to be more useful. Surely, geologists can continue to use the data of the engineering profession without being obligated to employ the specialized vocabulary that accompanies it.

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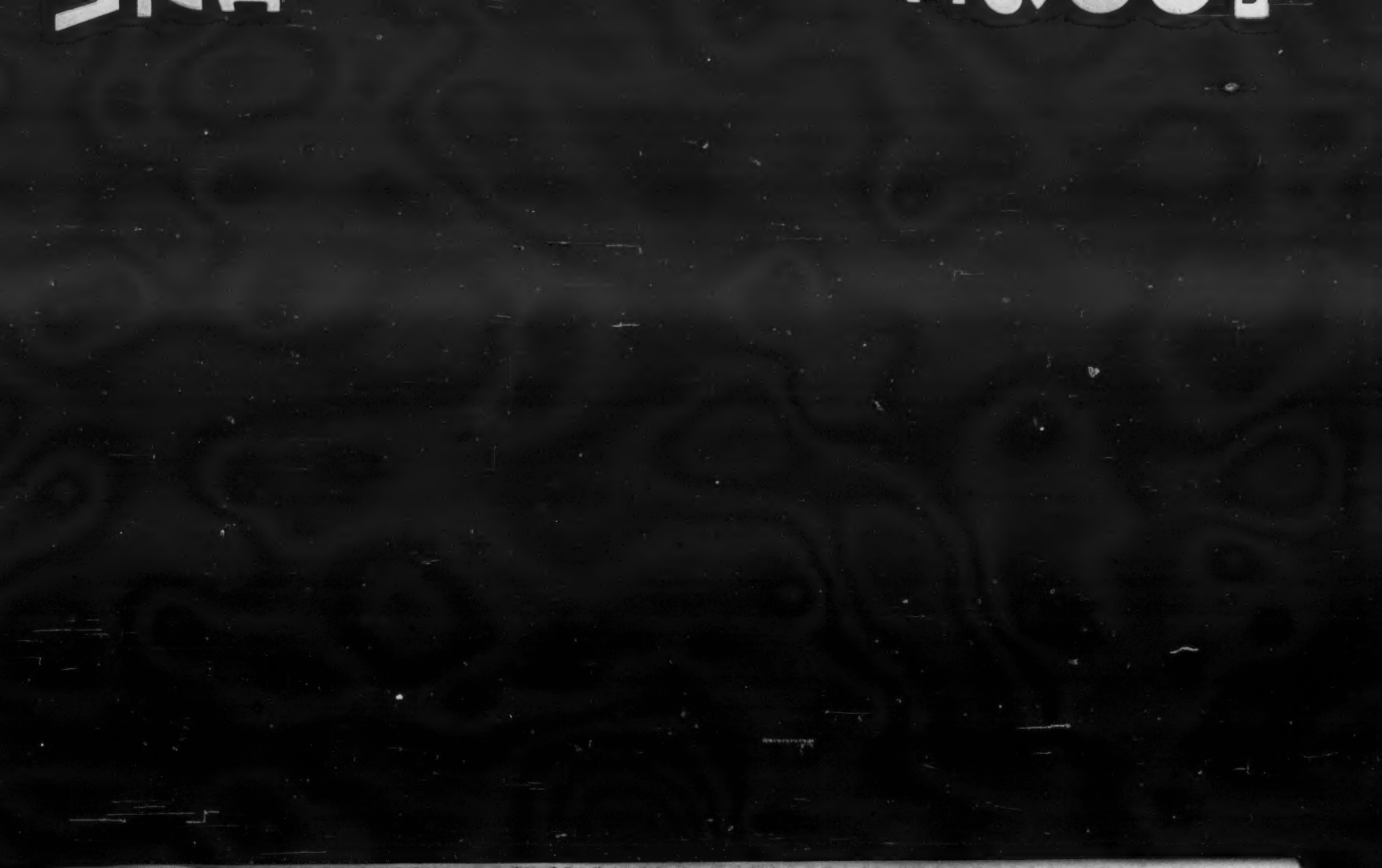

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